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## Origin and Significance of Olistostromes in the evolution of orogenic belts: a global syntesis

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# Origin and significance of olistostromes in the evolution of orogenic belts: A global synthesis

Andrea Festa [a,\\*](#), Kei Ogata [b,1](#), Gian Andrea Pini [c](#), Yildirim Dilek [d](#), Juan Luis Alonso [e](#)

a Dipartimento di Scienze della Terra, Università di Torino, 10125 Torino, Italy

b Dipartimento di Fisica e Scienze della Terra “Macedonio Melloni”, Università degli Studi di Parma, I-43124 Parma, Italy

c Dipartimento di Matematica e Geoscienze, Università di Trieste, 34128 Trieste, Italy

d Department of Geology and Environmental Earth Science, Miami University, Oxford, OH 45056, USA

e Departamento de Geología, Universidad de Oviedo, 33005 Oviedo, Spain

\* Corresponding author.

E-mail address: [andrea.festa@unito.it](mailto:andrea.festa@unito.it) (A. Festa).

1 Currently at Faculty of Earth and Life Sciences, VU University, 1081 HV, Amsterdam, The Netherlands.

## Abstract

We present a comparative analysis of the occurrences and internal structures of these sedimentary mélanges at a global scale with a focus on the Circum-Mediterranean, Appalachian and Circum-Pacific regions, and discuss their formation and time-progressive evolution in different tectonic settings. Lithological compositions, stratigraphy, and structural features of olistostromes reflect the operation of an entire spectrum of mass transport processes during their development through multi-stage deformation phases. The general physiography and tectonic setting of their depocenters, the nature, scale and rate of downslope transformation mechanisms, and global climatic events are the main factors controlling the internal structure and stratigraphy of olistostromes. Based on the tectonic settings of their formation olistostromes are classified as: (i) passive margin, (ii) convergent margin and subduction–accretion, and (iii) collisional and intra-collisional types. Systematic repetitions of these different olistostrome types in different orogenic belts provide excellent markers for the timing of various tectonic events during the Wilson cycle evolution of ocean basins. Olistostromes are best preserved in paleo active margins, covering vast areas of thousands of km<sup>2</sup>, where they underwent significant downslope translation, up to hundreds of kilometers. Incorporation of olistostromes into subduction–accretion complexes and orogenic belts takes place during discrete episodes of tectonic events, and their primary (sedimentary) fabric may be commonly reworked and overprinted by subsequent phases of tectonic and metamorphic events. We apply the basic nomenclature of structural geology, sedimentology and basin analysis in studying the internal structure, lithological makeup, and mechanisms of formation and extraordinary downslope mobility of olistostromes.

**Keywords:** Olistostrome, Mass transport deposits, Sedimentary mélange, Subduction–accretion complex, Orogenic belts

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## 1. Introduction

### 1.1. Objectives of this study

In this paper, we present an overview of the occurrences of olistostromal deposits in the Circum-Mediterranean, Appalachian and Circum-Pacific orogenic belts, and discuss their internal structures and origins through a comparative global synthesis. We define the significance of olistostromes in the tectonic evolution of orogenic belts and exhumed subduction–accretion complexes, and examine the mode, nature and interplay of different geological processes that commonly operate during development of these “sedimentary mélanges”. In the first part of the paper, we re-evaluate the spatial and temporal relationships between different types of olistostromes and the tectonic settings of their formation. Then, we show and discuss how the physical conditions of preservation may explain the apparent contrasts in the size and distribution of olistostromes in modern and ancient depositional settings. In the third part of the paper (Section 4), we examine and discuss the causes of observed variations in the size, internal fabric and run-out emplacement distance of olistostromes, while addressing their emplacement mechanisms. Following the work of Ogata et al. (2016), we also demonstrate that a proper application of the terminology of standard structural geology, in addition to using the sedimentological nomenclature, is highly useful in describing the internal fabric of olistostromes. The results of this global synthesis (Section 5) indicate that detailed, multidisciplinary studies of olistostromes are insightful for better understanding and constraining the temporal evolution of different stages of orogenic buildup and crustal exhumation in the Precambrian and Phanerozoic rock records, also pointing out their close correlation with global climatic events.

### 1.2. Definition of “olistostrome”

The term “olistostrome” (from the Greek “olistomai” — to slide and “stroma” — accumulation) was first introduced by Flores (1955) to define “sedimentary deposits occurring within normal geological sequences that are sufficiently continuous to be mappable, and that are characterized by lithologically and (or) petrographically heterogeneous materials, more or less intimately admixed, that were accumulated as a semi-fluid body”. Flores (1955, 1956) further specified that olistostromes do not show true internal bedding, and that they can be differentiated in a matrix (“binder”), which consists of “prevalent pelitic, heterogeneous material”, and dispersed “bodies of harder rocks” (“from pebbles up to several cubic km”). The introduction of the concept of “precursory olistostrome” by Elter and Trevisan (1973) played a fundamental role in better understanding the formation and evolution of collisional orogenic belts. This concept followed the wildflysch notion in the Alps presented earlier by Kaufmann (Kaufmann in Studer, 1872; Kaufmann, 1886; see Mutti et al., 2009 for a complete review), and defined the origin of gravitational chaotic deposits emplaced in front of advancing nappes. This conceptual development significantly increased our understanding of the relationships between tectonic and depositional processes in accretionary complexes and in fold-and-thrust belts. Subsequently, the term olistostrome has been used extensively by the international geological community in reference to stratally disrupted, “chaotic” complexes and “exotic” sedimentary packages, resulted from various mass transport events.

In this extended definition, the “olistostrome concept” played a major role in long-lived debates, representing the counterpart of tectonic mélanges (i.e., olistostromal mélange or sedimentary mélange; see, e.g., Hsü, 1974; Raymond, 1984; Bettelli and Panini, 1985; Cowan, 1985; Camerlenghi and Pini, 2009; Festa et al., 2010a, 2012b, 2014b). Specifically, the term “allolistostome” (Elter and Raggi, 1965), which was proposed to describe olistostromes containing both native (i.e., intraformational) and exotic (i.e., extraformational) blocks (Table 1), focuses on the significance of sedimentary processes for the genesis of mélanges (i.e. sedimentary mélanges). The term “endolistostrome” (Elter and Raggi, 1965), which describes olistostromes containing only native blocks (Table 1), represents the equivalent sedimentary product of a “broken formation” (e.g., Raymond, 1984). In most orogenic belts and exhumed subduction–accretion complexes, a strong morphological similitude of meso- to map-scale fabric elements exists between a block-in-matrix fabric of basin-wide olistostromes and a tectonic mélange. This resemblance remains at the base of a long-lasting debate on the nature and mode of geological processes that lead to the formation of chaotic rock assemblages

(i.e. gravitational vs. tectonics), particularly in those areas of well-preserved, exhumed subduction–accretion complexes, such as in the Western US Cordillera and in the Circum-Pacific Region (see, e.g., [Berkland et al., 1972](#); [Aalto, 1981](#); [Cloos, 1982](#); [Cloos, 1984](#); [Raymond, 1984](#); [Cowan, 1985](#); [Brandon, 1989](#); [Ukar, 2012](#); [Wakabayashi, 2012](#); [Aalto, 2014](#); [Platt, 2015](#); [Raymond and Bero, 2015](#); [Ukar and Cloos, 2015](#); [Wakabayashi, 2015](#); [Ukar and Cloos, 2016](#)). The main aspect of the debate (see [Alonso et al., 2006, 2015](#)) is whether the “chaotic disruption” of rock assemblages beneath long-traveled nappe systems is a result of thrust related shearing (“tectonic brecciation”, see, e.g., [Bailey and McCallien, 1950](#); [Vollmer and Bosworth, 1984](#); [Burkhard, 1988](#); [Jeanbourquin, 1994](#)), or a product of mass wasting, which causes denudation of a nappe toe (e.g., [Page, 1962](#); [Boccaletti et al., 1966](#); [Debelmas and Kerchkov, 1973](#); [Elter and Trevisan, 1973](#); [Aalto, 1981](#); [Behr et al., 1982](#); [Yilmaz and Maxwell, 1984](#); [Alonso et al., 2006](#); [Codegone et al., 2012a](#); [Alonso et al., 2015](#)).

### 1.3. Olistostromes and mass transport deposits

A vast amount of geophysical data acquired from modern continental margins show that mass transport deposits (MTDs) and complexes (MTCs) exist at various scales, and with varying abundance and characteristics, as controlled by the mode, nature and interplay of different geological processes in their tectonic setting of formation (e.g. [Mosher et al., 2010](#); [Yamada et al., 2012](#); [Krastel et al., 2014](#); [Lamarche et al., 2016](#); [Moscardelli and Woods, 2016](#) and reference therein). These mass transport deposits represent either the products of a single depositional event or composite bodies originated by superposed, multiple events (e.g., [Weimer, 1989](#); [Weimer and Shipp, 2004](#); [Della Valle et al., 2015](#)), which may involve sediments with different degree of consolidation/lithification and grain-size (from clay to silt, to sand, to gravel size; [Elverhøi et al., 2010](#); [Strasser et al., 2011, 2012](#)). MTDs and MTCs are commonly characterized by great internal heterogeneity and deformation, resulting in acoustic artifacts and transparent zones in 2D and 3D seismic imagery ([Gamboa et al., 2010](#); [Strasser et al., 2011, 2012](#); [Ogata et al., 2014a](#); [Alves, 2015](#)). However, an increasing amount of studies based on marine geophysical data and core analysis show a partition of internal structural arrangement into discrete deformation domains, suggesting: (i) differential movement of discrete bodies of mass during translation and emplacement, (ii) episodic pulses during the same depositional event(s) and, (iii) interplay of different, synchronous mass transport processes ([Vanneste et al., 2011](#); [King et al., 2011](#); [Ogata et al., 2012a](#); [Omosanya and Alves, 2013](#); [Joanne et al., 2013](#)).

Sedimentary rock bodies that formed as a result of mass transport processes widely occur in orogenic belts and in exhumed subduction–accretion complexes ([Fig. 1](#)), representing the ancient, “fossil” counterparts of modern MTDs and MTCs ([Lucente and Pini, 2008](#); [Ogata et al., 2012a](#)). The term olistostrome that was first introduced by [Flores \(1955\)](#) to describe “chaotic” mass transport deposits with a shaly to pelitic, brecciated matrix in the Apennines (see [Ogniben, 1953](#); [Rigo de Righi, 1956](#)) has been used extensively in the literature. Its application has been also widened beyond the original definition to encompass chaotic rock bodies with various matrix types, including sandy/silty (e.g., [Abbate et al., 1970](#); [Hsü, 1974](#); [Erickson, 2011](#); [Okay et al., 2012](#)), volcanic (e.g., [Kovacs, 1989](#)), ophiolitic (e.g., [Abbate et al., 1970](#); [Gansser, 1974](#); [Tekeli and Erendil, 1986](#); [Lagabrielle, 1994](#)), clast-sustained debrites ([De Libero, 1998](#)), single blocks (olistoliths) or aggregates of slide blocks (Type C olistostromes of [Lucente and Pini, 2003](#)).

Olistostromes are in general composite deposits displaying superposed structures and complex stratigraphic relationships that developed during multiple events and mass transport processes (sliding, slumping, debris flow, blocky flow, turbidity currents; see, e.g., [Lucente and Pini, 2003, 2008](#); [Pini et al., 2004](#); [Festa et al., 2013, 2015c](#)). Downslope motion of olistostromal deposits is enabled by the relative movement of discrete bodies of masses, causing progressive strata disruption and slide-flow transformation (see, e.g., [Lucente and Pini, 2003](#)).

The extended use of the term olistostrome, which we follow in this study, can be applied to different types of fossil MTDs and MTCs, characterized by disrupted to “chaotic” internal fabrics, regardless of its original definition, which simply implies a genetic significance. Even with this broader definition, this term should not be used to indicate the parent sedimentological process, since it refers to the depositional product (i.e. deposit, accumulation). The existing inherent literature already points out a largely accepted terminology (e.g., block flow, debris flow, see [Mutti et al., 2006](#); [Ogata et al., 2012b](#); [Shanmugam, 2015](#)). Although this historical term clearly indicates “chaotic” sedimentary bodies, its use is transversal to both sedimentologists and structural geologists, having a strong tectonic and geodynamic significance.

## 2. Olistostromes and tectonic settings of their formation

The Circum-Mediterranean, Appalachian, and Circum-Pacific regions include some of the most extensively studied orogenic belts and exhumed subduction–accretion complexes, which display well preserved olistostromes of different ages ([Fig. 2](#); [Dilek, 2006](#)). On the basis of a comparative analysis of exhumed, fossil olistostromes, and following the earlier work of [Festa et al. \(2010a, 2010b\)](#); see also [Camerlenghi and Pini, 2009](#); [Festa et al., 2012b](#)), we show that close relationships exist between the types of olistostromes and the tectonic settings of their formation ([Table 2](#)). Each type of olistostrome is also distinguished in terms of its composition and internal fabric, according to the source area and mechanism of emplacement.

### 2.1. Passive margin tectonic settings and olistostromes

Extensional tectonics and rifting-related geological processes commonly lead to the formation of various types of olistostromes during the course of passive margin evolution. Such olistostromes develop at the edges of thinned continental margins and carbonate platforms, at ocean–continent transition zones (OCT), and along oceanic core complexes (Table 2 and Fig. 3).

Along platform margins (Fig. 3C), olistostromes are generally the product of en-masse gravitational movements (e.g. debris flow, debris avalanches, and block sliding), and comprise megabreccias, olistoliths and olistolith fields (Table 2). They are commonly characterized by angular clasts and blocks (decimeters to several tens of meters in size; Fig. 3I, L) and subordinate and smaller, rounded clasts, which are older than the surrounding, fine-grained matrix. Such olistostromal material is mainly sourced from intrabasinal topographic highs (e.g. horsts) and basin margin shoulders, all bounded by master normal faults. It can also be derived from tectonically uplifted, extended carbonate platforms along rifted continental edges (Fig. 3C, I, L). The matrix is predominantly made of pelagic limestone. Some of the best examples of this type of olistostromes exist in the Southern Alps and in the Northern Calcareous Alps (e.g., [Castellarin, 1972](#); [Channell et al., 1992](#); [Böhm et al., 1995](#); [Ortner, 2001](#); [Amerman et al., 2009](#)), Apennines (e.g., [Bernoulli, 2001](#); [Graziano, 2001](#)), Western Hellenides ([Naylor and Harle, 1976](#); [Ghikas et al., 2010](#)), and Appalachians (e.g., [Rast and Kohles, 1986](#); [Bailey et al., 1989](#)).

The most representative olistostrome types at OCTs (Fig. 3B) are poorly sorted ones with blocks commonly consisting of finegrained carbonate, siliciclastic turbidites, and/or chaotic brecciated (i.e., matrix-supported) masses (Fig. 3G, H). Such lithological units are monomictic (i.e., dominated by native/intrabasinal clasts) when formed adjacent to rifted passive margins, and polymictic (i.e., mixing of native and exotic clasts) when developed proximal to an oceanic realm (Table 2), where the matrix material consists of deep-sea sediments. Polygenetic blocks and olistoliths may range from tens of meters to several kilometers in size. Hydroplastic deformation of blocks and clasts, liquefaction/fluidization, and softsediment deformation of the matrix (e.g., fluidal features, in situ folding, boudinage) indicate that sediments were non- to poorlyconsolidated at the time of olistostrome formation.

Reactivation of pre-existing extensional faults, fluctuations in sealevel and dissociation of gas-hydrates are the most common triggering mechanisms influencing slope instability, which is closely related to the shear strength of sediments. Olistostromes that formed in paleo OCTs are widely documented (Fig. 2) from the Circum-Mediterranean Region (e.g., Apennines, see [Naylor, 1982](#); [De Libero, 1998](#); [Pini et al., 2004](#); Hellenides–Albanides, see [Smith et al., 1979](#); [Shallo, 1990](#); [Shallo and Dilek, 2003](#); Taurides, see [Dilek and Rowland, 1993](#)), the Central Appalachians (e.g., [Jacobi and Mitchell, 2002](#); [Wise and Ganis, 2009](#); [Codegone et al., 2012a](#)), and the Argentine Precordillera ([Banchig, 1995](#); [Keller, 1999](#); [Alonso et al., 2008](#)). The best examples of analog MTDs occur in modern settings worldwide (e.g. Storegga, Saharan; [Bugge et al., 1988](#); [McAdoo et al., 2000](#); [Mienert et al., 2003](#), and references therein).

In an oceanic realm (Fig. 3A), collapse of topographic highs of upper mantle abyssal peridotites, associated with mid-oceanic ridge settings and seamounts mainly form debris flows (e.g., [Dilek and Rowland, 1993](#); [Sarifakioglu et al., 2014](#); [Liu et al., 2015](#)) commonly consisting of clast- to matrix-supported angular clasts of mafic- to ultramafic-rocks, embedded in a matrix composed mainly of pelagic limestone and/or medium- to coarse-grained sandstone with ophiolite-derived detrital material (Fig. 3E–F). Ophiolitic olistoliths and olistolith fields, related to block sliding and debris avalanche also occur close to mantle topographic highs and seamounts. Well-documented examples of olistostromes that developed at ancient mid-ocean ridge settings occur in the Western Alps and Pyrenees (e.g., [Lagabriele et al., 1984](#); [Lagabriele, 1994](#); [Lagabriele and Lemoine, 1997](#); [Clerc et al., 2012](#); [Balestro et al., 2014, 2015a, 2015b](#); [Festa et al., 2015a](#); [Tartarotti et al., 2015](#)), in the Apennines (e.g., [Abbate et al., 1970](#); [Decandia and Elter, 1972](#); [Bortolotti et al., 2001](#)), and in the Western US-Cordillera (e.g., [Saleeby, 1979](#)).

## **2.2. Convergent margin and subduction zone environments and olistostromes**

Olistostromes forming in these settings (Table 2 and Fig. 4), including fore-arc basins, are generally characterized by different degrees of stratal disruption, related to the consolidation state at the time of the slope failure and the final run-out distance of slide masses. Olistostromal material includes deformed sediments and native, extrabasinal or exotic rocks of different ages, generally older and more consolidated than intrabasinal components (Fig. 4E, F, G, I), which are sourced from the accretionary wedge front and/or wedge-top basins. Extrabasinal clasts and blocks comprise chunks or entire packages of sedimentary beds, which locally display their subduction-related tectonic fabric elements. The matrix varies from shale and generally fine-grained sediments (Fig. 4E, I) to medium- to coarse-grained sandstones (Fig. 4F, G). Pinch-and-swell structures due to boudinage development, slump folds and “slump balls” are the most common features in the matrix, together with meso-scale contractional and extensional duplexes, imbricated bedding, isoclinal folds, and shear zones (e.g., [Taira et al., 1992](#); [Yamamoto et al., 2009](#); [Ogata et al., 2016](#)). Debris flow, debris avalanches, sliding and slumping are the main modes of mass-transport processes that interact and overlap with each other during the formation of this type of olistostrome.

Tectonic over-steepening is the most important process, among several other factors, controlling slope instability at accretionary wedge fronts and retro-wedge fronts of doubly-verging accretionary wedges (e.g., [Malavieille et al., in press](#)). This process is generally associated with “basal” and “frontal tectonic erosion” (sensu [Von Huene and Lallemand, 1990](#)), subduction erosion, seamount and ridge subduction ([Anma et al., 2011](#); [Kawamura et al., 2011](#)), and thrust faulting and folding (see [Martinez Catalan et al., 1997](#); [Marroni and Pandolfi, 2001](#); [Von Huene et al., 2004](#); [Ruh, 2016](#)).



Upward rise of overpressurized fluids and fluidized/liqefied unconsolidated sediments sourced from out-of-sequence thrust faults and/or basal detachments of the frontal- and/or retro-wedges or subduction interface represents another possible triggering mechanism for slope failure (e.g., Barber et al., 1986; Lash, 1987; Conti and Fontana, 2011; Codegone et al., 2012b; Barber, 2013; Festa et al., 2013, 2015c). Some of the most important examples of subduction-related olistostromes occur in the Circum-Mediterranean orogenic belts, such as the Apennines (see Abbate et al., 1970; Elter and Trevisan, 1973; Bertotti et al., 1986; Pini, 1999; Marroni and Pandolfi, 2001; Malavieille et al., in press), the Corsican and Western Alps (see Polino, 1984; Durand Delga, 1986; Deville et al., 1992), the Central Anatolian range (see Dangerfield et al., 2011; Sarifakioglu et al., 2014), the Mamonia Complex in SW Cyprus (see Swarbrick and Naylor, 1980), in the Appalachians (see, e.g., Lash, 1987; Wise and Ganis, 2009; Codegone et al., 2012a), and in the Circum-Pacific orogenic belts (e.g., Western US Cordillera, see Aalto, 1981, 2014; Raymond and Bero, 2015; Wakabayashi, 2015; Dominican Republic, see Hernaiz Huerta et al., 2012; New Zealand, see Chanier and Ferrière, 1991; Delteil et al., 2006; Japan, see Aoya et al., 2006; Yamamoto et al., 2009; Osozawa et al., 2011). This type of olistostrome has been also reported from fore-arc basins (Fig. 4D) at the rear of accretionary wedges (see Hitz and Wakabayashi, 2012 for a US-Western Cordillera example; Escuder-Viruete et al., 2015 for Hispaniola in the Caribbean Plate; Lamarche et al., 2008 for the modern Kermadec fore-arc basin in NE New Zealand). In the Circum Mediterranean Region, Late Cretaceous fore-arc related olistostromes were derived from the internal side of the Alpine accretionary wedge (i.e., present-day Northern Apennines) and were then emplaced in a supra-subduction basin (Bertotti et al., 1986; Elter et al., 1991).

Emplacement of sections of lithospheric mantle onto continental margins represents another process, by which convergent margin related olistostromes develop. Some of the best examples include the Al Hajar Mountains—Oman (e.g., Michard et al., 1991), Albanides (e.g., Bortolotti et al., 1996; Dilek et al., 2005), Hellenides (e.g., Jones and Robertson, 1991; Bortolotti et al., 2003; Ghikas et al., 2010), Western Anatolides (e.g., Okay et al., 2012), and the Coastal New Zealand (Delteil et al., 2006). Emplacement-related olistostromes become accreted (Fig. 4D) as a section of subducting oceanic lithosphere is thrust into the accretionary wedge as accretionary type ophiolites (Dilek and Furnes, 2011, 2014). Blocks derived from both oceanic and continental crustal assemblages characterize this type of olistostrome. Oceanic material is sourced from the front of an obducting ophiolite sheet (i.e., obduction-related olistostromes), and continental material from the continental margin succession beneath the advancing ophiolite nappe (Bortolotti et al., 2013). Debris flows consisting only of ophiolitic blocks and associated sedimentary rocks may also be emplaced during an intra-oceanic obduction stage (Lagabrielle et al., 1986) when a fragment of oceanic lithosphere tectonically overrides another oceanic plate.

### 2.3. Collisional and intra-collisional tectonic settings and olistostromes

These types of olistostromes (Table 2; Fig. 4B, C), particularly the intra-collisional ones, represent the majority of examples documented from ancient orogenic belts in the Circum-Mediterranean Region and from the Appalachians, and are directly related to the early phases of mountain-building processes. Festa et al. (2010a, 2012b; see also Camerlenghi and Pini, 2009) subdivided these types of olistostromes into three main sub-types (i.e., sub-nappe, intra-nappe and epinappe) on the basis of their relative position with respect to the allochthonous nappe structures.

Sub-nappe olistostromes are, in turn, subdivided into: (i) precursory olistostromes, and (ii) olistostromal carpet (Table 2 and Fig. 4C). The precursory olistostromes (sensu Elter and Trevisan, 1973) consist of classic olistostromes (and/or wildflysch; e.g. Mutti et al., 2009) with a block-in-matrix fabric, emplaced by cohesive debris flows (Fig. 4L) and/or block avalanches in migrating foredeep basins (e.g., Bird, 1969; Root and MacLachlan, 1978; Behr et al., 1982; Frisch, 1984; Pini, 1999; Lucente and Pini, 2003, 2008; Masson et al., 2008; Vezzani et al., 2010; Festa et al., 2010b, 2012b; González Clavijo et al., 2016). This type of olistostrome is an end product of a “closure event”, recording the depositional deactivation on top of the foredeep turbidite successions. It forms before the advancement of thrust-related deformation and subsequent incorporation into a collisional belt. Some of the excellent examples include the Aveto and Macigno Formations in the Northern Apennines (e.g. Lucente and Pini, 2003, 2008), and the Tarakli Flysch in Turkey (e.g., Catanzariti et al., 2013). Olistostromal carpets (Pini et al., 2004) comprise laterally and vertically coalescing and overlapping aprons of failed sediments, and develop as a result of protracted debris flow activities (Fig. 4C) and avalanches in front of advancing nappe sheets. Such avalanche deposits are then structurally overridden by allochthonous nappes during their tectonic migration. Blocks and discrete slides are sourced from nappe fronts (Fig. 4M), whereas loose to poorly-consolidated sediments originate from thrust-top and intraslope basins and slope deposits emplaced atop the nappe front (Alonso et al., 2006).

In this framework superposition of tectonic shearing (Fig. 4N) exerted by nappe emplacement on primary gravitational fabric elements of olistostromes typically complicates the final products. Some of the salient examples have been documented, for example, from the base of the Taconic thrust front in the Appalachians (see, e.g., the Huston River mélange of Vollmer and Bosworth, 1984; Bosworth, 1989; the Poetstenkill Gorge mélange of Festa et al., 2012a; some mélanges in the Hamburg Klippe of Lash, 1987; Codegone et al., 2012a), in the Central Alps (Habkern mélange of Kempf and Pfiffner, 2004), some mélanges at the base of the Ligurian Units in the Northern Apennines (e.g., Lucente and Pini, 2008; Ogata et al., 2012a; Festa et al., 2013) and in the Southern Apennines (e.g., Mattioni et al., 2006; Festa et al., 2010b; Vezzani et al., 2010), in the Anatolide–Tauride orogenic belts (e.g., Bailey and McCallien, 1950, 1953; Dilek and Delaloye, 1992; Dilek et al., 1999; Sarifakioglu et al., 2012), Othris Mountain in Greece (Smith et al., 1979), and in Taiwan (Page and Suppe, 1981).

Intra-nappe olistostromes (Table 2), which develop as a result of tectonic deformation during nappe translation, mainly consist of blocks of intrabasinal origin and/or portions of older sedimentary successions embedded within a matrix made of the same lithologies. Conglomerates, breccias–megabreccias, and large isolated olistoliths can be the products of both rock-fall and gravity flow processes. Large, intra-nappe shear zones, associated with out-of-sequence thrusting, may form olistostromal carpet-like intervals, only differing from the typical ones in their position within the nappe (Festa et al., 2010a and reference therein).

Epi-nappe olistostromes (Table 2 and Fig. 4C), which form by gravitational instability along the margins of piggyback, thrust-top, wedge-top, episutural and satellite basins (Fig. 4O), represent the most common types in the Circum-Mediterranean Region. Both blocks and the matrix in these olistostromes are sourced from the stratigraphic successions that are tectonically dismembered and imbricated in thrust sheet stacks (e.g., Papani, 1963; Bettelli and Panini, 1989; Bettelli et al., 1989, 1994; Pini, 1999; Panini et al., 2002; Ferrière et al., 2004; Festa et al., 2005; Remitti et al., 2011; Ogata et al., 2012b; Festa and Codegone, 2013; Ogata et al., 2014a; Festa et al., 2015b, 2015c).

Triggering mechanisms for gravitational instability in collisional and intra-collisional settings are mainly seismic shocks emanating from thrust faulting. However, in epi-nappe settings climatic forcing may also play a significant role, primarily by enhancing the sediment supply and by causing important variations in sea-level and subsidence rates.

### 3. Preservation of olistostromes in the geological record

Relationships between different types of olistostromes and their tectonic settings of formation, as described above, are directly influenced by their preservation potential in the geological record. In contrast to modern tectonic settings, in which the majority of such deposits occur in association with gravitational instability along passive margins (e.g., Macdonald et al., 1993; Mienert et al., 2003; Camerlenghi and Pini, 2009), olistostromes originated from active margins are more widespread in ancient orogenic belts and exhumed subduction–accretion complexes (Fig. 5). Moreover, submarine landslides, described on modern continental margins, are documented to be several orders of magnitude larger in cross-sectional area than their equivalent olistostromes (e.g., Woodcock, 1979).

In the following, we provide independent lines of evidence to discuss and reevaluate these existing apparent dichotomies, taking into consideration that olistostromes may represent either the product of a single depositional event (i.e., MTD) or composite bodies (i.e., MTC), the origin of which is due to superposed events (Heck and Speed, 1987; Macdonald et al., 1993; Lucente and Pini, 2003; Pini et al., 2012; Ogata et al., 2012a, 2012b, 2014a). Three main factors effectively influence the preservation of olistostromes: (i) their transformation during transfer and emplacement, (ii) their size, and (iii) the role played by the overprinting of tectonic deformation during subsequent orogenic stages.

#### 3.1. Transformation during emplacement: olistostromes as ancient examples of mass-transport complexes

Olistostromes, particularly large ones, are complex rock assemblages, whose formation may involve the entire spectrum of mass transport processes (Lucente and Pini, 2003), and they comprise discrete sub-units, subdivided on the basis of composition, provenance, structures, and sense of movement (Figs. 6 and 7). These sub-units are interpreted as products of discrete slide masses moving differentially, and of multiple, overlapping submarine en-masse flow processes (Ogata et al., 2014b; see also Strachan, 2002; Lucente and Pini, 2003; Ogata et al., 2012b). The final anatomy of the olistostrome, as a slide body, depends on the extent of segregation/separation among the different internal structural facies associations (Fig. 7). Analogous to the concept of “flow efficiency” for turbidites as introduced by Mutti et al. (1999), olistostromes characterized by high efficiency may, for example, separate different facies populations horizontally during their downslope transport, hence forming deposits with laterally varying facies associations and marked boundaries (Fig. 7). Such deposits are expected in passive margins, where continental slopes are characterized by relatively smooth morphologies and constant gradients. On the other hand, low efficiency slides appear as composite bodies characterized by strong lateral–vertical variations, horizontal heteropy, and interfingering of different facies associations as a result of their incomplete/partial segregation. These types of deposits likely develop in depositional settings having a complex physiography with structurally confined depocenters separated by intrabasinal highs. Such features constitute local morphological barriers and hence favor flow transformations on short distances.

Recent studies (Lucente and Pini, 2003, 2008; Pini et al., 2012; Ogata et al., 2012b) have shown that basin-wide olistostromes or mass-transport complexes (MTC) in the Northern Apennines consist of different structural associations: (i) mud-rich chaotic deposits characterized by the cohesive-viscous behavior of a viscous matrix (Type 1 MTC; see Fig. 7(3)), including cm- to m-sized lithic blocks randomly distributed in a brecciated or clastic matrix. The matrix may sustain bedding packages that are tens of meters and up to kilometers in size (floaters); (ii) clastic matrix represented by an unsorted, liquefied/fluidized mixture of different grain-size populations, dispersed within a fine-grained lithology (Type 2 MTC). Such matrix material may show the characteristic features of a hyper-concentrated suspension (sensu Mutti, 1992; see also Callot et al., 2008; Ogata et al., 2012b, 2014a), commonly arranged as bipartite bodies with a block dominated portion overlying a matrix-dominated one. This kind of matrix may originate from a complex debris flow carrying coherent blocks (meters- to hundreds of meters wide), which are commonly arranged in isolated slump-like folds (blocky-flow deposits sensu Mutti et al., 2006; see Fig. 7(2)); (iii) slump-slide-like MTC (Type 3 MTC; see Fig. 7(1)) developed in sandy sediments (e.g. foredeep turbidites), in which transport and emplacement are allowed by differential movement of individual bed-packages of different dimensions. The relative movement is accommodated



along ductile shear zones, which are marked by mm-thick “films” of unsorted silty material and cm- to tens of cm-thick deformation/disaggregation bands in sandstones. These shear zones develop along with transportation/rotation of clasts and elutriation of finer-grained clastic material populations (see also [Dykstra, 2005](#)).

The Miocene Casaglia–Monte della Colonna body (N350 km<sup>2</sup> and up to 300 m thick; see [Lucente and Pini, 2003, 2008](#)), the Oligocene Specchio Unit (about 1500 km<sup>2</sup> and average thickness of 100 m; see [Ogata et al., 2012c, 2014a](#)), the late Oligocene–early Miocene Val Tiepido–Canossa argillaceous breccias (over an area of about 300 km long and tens of kilometers wide, and up to 300 m thick; see [Bettelli and Panini, 1985](#); [Panini et al., 2002](#); [Remitti et al., 2011](#); [Festa et al., 2015c](#) and reference therein) in the Northern Apennines, the Paleogene carbonate “megabreccia” (23 megabeds up to 1500 km<sup>2</sup> and up to 260 m thick each) of the Friuli Basin in the Southern Alps ([Ogata et al., 2014b](#)), and the Eocene foredeep carbonate megabreccias in the south-central Pyrenees ([Payros et al., 1999](#); [Ogata et al., 2012b](#)), represent some of the best examples of basin-wide olistostromes. These types of olistostromes display more than one type or all the structural associations described above.

### 3.2. The size paradox

[Woodcock \(1979\)](#) has pointed out earlier a size dichotomy between ancient submarine slide deposits and olistostromes observed in outcrops and those identified in modern continental margins ([Fig. 8](#); see also [Macdonald et al., 1993](#); [Lucente and Pini, 2003](#); [Camerlenghi and Pini, 2009](#); [Pini et al., 2012](#); [Ogata et al., 2014b](#)). Modern submarine slides appear to be, on average, several orders of magnitude larger in cross-sectional area than their inferred ancient counterparts ([Fig. 8](#)). Thus, the questions remain as to whether the upper limit of the size of ancient on-land olistostromes is a function of exposure and/or whether it reflects a real change in the magnitude of processes responsible for their formation (see [Woodcock, 1979](#); [Macdonald et al., 1993](#); [Pini et al., 2012](#)).

We can approach this size paradox to the first order by recognizing that olistostromes largely represent mass-transport complexes ([Pini et al., 2012](#)), showing substantial similarities in size ([Fig. 8](#)), distribution, recurrence interval (see below), and run-out distance ([Fig. 9](#)) with modern submarine slide deposits in different tectonic environments (e.g., [Camerlenghi and Pini, 2009](#); [Festa et al., 2010a](#); [Pini et al., 2010, 2012](#); [Ogata et al., 2014a](#)). For example, the Makran olistostrome, which is associated with an accretionary complex, represents one of the largest and best-preserved examples of an ancient sedimentary chaotic rock unit yet observed (about 42,000 km<sup>3</sup> and 10,000 km<sup>2</sup> distributed over an area of 72,000 km<sup>2</sup>, [Burg et al., 2008](#)). Direct comparisons are possible in similar tectonic and depositional settings, and the best examples include the Oligocene wedge-top basins of the Northern Apennines (e.g., [Bettelli et al., 1989](#); [Pini, 1999](#); [Remitti et al., 2011](#); [Pini et al., 2012](#); [Ogata et al., 2012c, 2014a](#); [Festa et al., 2015c](#)) and the modern Kumano Basin in the Nankai forearc setting (e.g., [Strasser et al., 2011](#)), the Poverty Unit in the Hikurangi margin of New Zealand (e.g., [Mountjoy and Micallef, 2012](#); [Ogata et al., 2014a](#)), and the MTCs from the Southern Magdalena Fan, offshore Colombia ([Ortiz-Karpf et al., in press](#)), the Late Cretaceous Casanova Complex of the Northern Apennines (e.g., [Bertotti et al., 1986](#); [Elter et al., 1991](#)) and the Matakaoa submarine instability complex in the Kermadec fore-arc basin in New Zealand ([Lamarche et al., 2008](#)). Some other examples also include the Miocene Makran olistostrome ([Burg et al., 2008](#)), the modern accretionary wedge of the Sunda Arc ([Moore et al., 1976](#)) and the Gibraltar Arc–Gulf of Cadiz ([Medialdea et al., 2004](#)), the Late Carboniferous Porma mélange and other mélanges of the Cantabrian Zone ([Alonso et al., 2006](#); [Martín-Merino et al., 2014](#); [Alonso et al., 2015](#)), the Ordovician mélanges of the Argentine Precordillera ([Heim, 1948](#); [Banchig, 1995](#); [Von Gosen et al., 1995](#); [Alonso et al., 2008](#); [Sobiesiak et al., 2016](#)), the Paleogene carbonate megabreccia units in NW Italy and Slovenia ([Ogata et al., 2014b](#)), the Early Miocene Nataraja submarine slide ([Calvès et al., 2015](#)) and the modern Giant Chaotic body offshore of Gibraltar ([Torelli et al., 1997](#)).

### 3.3. Sheared olistostromes or tectonic mélanges? A long-standing debate

In ancient orogenic belts and exhumed subduction–accretion complexes with a record of multiple deformational events, olistostrome fabrics are commonly overprinted and significantly reworked by tectonics ([Figs. 10D–E, and 11](#)), extremely complicating their distinction from tectonic mélanges (e.g., [Raymond, 1984](#); [Cowan, 1985](#); [Pini, 1999](#); [Festa et al., 2013](#) and reference therein). This is the case, for example, for olistostromal carpets formed at the base of intra-collisional nappes, and for olistostromes overprinted by thrust faulting and tectonic shearing in subduction zones and accretionary complexes. This problem, which remains at the core of a long-standing debate, especially in the Western US Cordillera (e.g., [Berkland et al., 1972](#); [Silver and Beutner, 1980](#); [Cloos, 1982](#); [Raymond, 1984](#); [Cowan, 1985](#); [Brandon, 1989](#); [Platt, 2015](#); [Raymond and Bero, 2015](#); [Ukar and Cloos, 2015](#); [Wakabayashi, 2015](#) and reference therein) and in all exhumed subduction–accretion complexes throughout the world (e.g., [Pini, 1999](#); [Şengör, 2003](#); [Alonso et al., 2006, 2008](#); [Festa et al., 2010a, 2012b](#); [Alonso et al., 2015](#); [Wakita, 2015](#); [Balestro et al., 2015b](#)), revolves around the mechanisms of incorporation and mixing of exotic blocks within a matrix. Although the “flow mélange model” ([Cloos, 1982, 1984](#)) is the most important mixing process at deeper structural levels within subduction channels ([Fig. 10B, C](#)), gravitational processes represent the most effective ways in incorporating exotic blocks into a shale matrix at shallow structural levels (e.g., [Cowan, 1985](#); [Alonso et al., 2006](#); [Festa et al., 2012b](#); [Platt, 2015](#)). Exotic blocks, which are supplied by mass transport events, for example, at the front of a nappe ([Fig. 10A–C](#)), can be easily misinterpreted as blocks incorporated through tectonic slicing and shearing, and vice versa, because the two end products (i.e., olistostromes overprinted by tectonics and tectonic mélanges) show strong morphological similarities of rocks with similar block-in-matrix fabrics ([Figs. 10D–E, 11](#)). The composition of the matrix, however, may provide

significant information on the tectonic setting of formation (see [Section 2](#)). For example, the occurrence of sandstone-matrix mélanges, rather than shale matrix ones, indicates that gravitational processes, which formed olistostromes, contributed to the formation of at least part of the well-known Franciscan Complex (e.g., [Aalto, 1981, 2014](#); [Raymond and Bero, 2015](#) and reference therein).

Olistostromes with bedded “floaters”, such as slide blocks composed of specific turbidite facies and grain-flow deposits, embedded in a sandstone matrix, are stratigraphically and sedimentologically consistent with parts of lower slope-base of slope sedimentary successions, arguing against flow mélange origins in a subduction channel ([Raymond and Bero, 2015](#)). In this framework, the compositional attributes of matrix and blocks may provide significant clues to better understand the evolution of subduction complexes and the structural architecture of the frontal part of accretionary wedges. Accumulation of thick and sandy (and/or heterogeneous) sediments within an accretionary complex (Figs. 10B, C, 11) may result in frontal accumulation of thick wedges of sandstone-dominated mass transport complexes (e.g., [Seely et al., 1974](#); [Ernst, 1977](#); [Clift and Vannucchi, 2004](#); [Raymond and Bero, 2015](#)), rather than thinner, shale-matrix fault zone mélanges, which may form and accrete incrementally (e.g., [Cowan, 1985](#); [Festa et al., 2010a](#); [Rowe et al., 2013](#)). This, in turn, may suggest that in some cases a significant reassessment of transfer mechanisms and amounts of material transferred from the overriding plate to the subducting plate is required (see [Raymond and Bero, 2015](#) for a detailed discussion). Submarine mass transport complexes and olistostromes, occurring in frontal parts of accretionary wedges, may locally reach volumes up to thousands of km<sup>3</sup> in size (e.g., [Macdonald et al., 1993](#); [Hampton et al., 1996](#); [Collot et al., 2001](#); [Lewis et al., 2004](#); [Burg et al., 2008](#); [Contreras-Reyes et al., 2016](#); [Ruh, 2016](#); see [Section 3.2](#) and [Fig. 10A](#)). Thus, they play a crucial role in controlling the thickness of accretionary wedges, their internal architecture and lateral morphological variations. Different trajectories may be followed ([Fig. 10B, C](#)) in accretionary versus erosional margins.

#### **4. Mechanisms of olistostrome emplacement: interplay between structural geology and sedimentology**

A systematic examination of basin-wide olistostromes indicates unusually fast-rated processes and long-distance mobility (tens to hundreds of kilometers) of thick (200–300 m) cohesive olistostromes ([Fig. 9](#)). This observation contrasts with their internal fabric and composition and with the fact that they do not fully transform during emplacement. Cohesive olistostromes (e.g., Type 1 MTD of [Pini et al., 2012](#) should be deposited by relatively slow-moving slide masses, in which the movement is achieved through laminar “viscous” shear zones in a clay-dominated matrix ([Fig. 12](#)). Thus, thick cohesive olistostromes are not expected to have high mobility and long run-out into a basin.

##### **4.1. Large-scale olistostromes: a paradox in transition, internal fabric and emplacement?**

Detailed field observations have shown that the emplacement of some cohesive olistostromes, such as clay-rich cohesive debris flows, may have been a fast-paced process capable of carrying material for long distances when sustained by a thin (up to a meter in thickness) and continuous, basal shear zone. Much of deformation takes place along this basal shear zone, which represents a hydroplaning “carpet” ([Fig. 12B](#)) consisting of a liquefied/fluidized mixture of water and loose sediments (e.g., [Pini et al., 2012](#); [Ogata et al., 2014b](#); [Festa et al., 2015c](#)). This inference is consistent with observations from both modern and submarine landslides, and laboratory experiments (e.g., [De Blasio et al., 2006](#) and reference therein). The process of “autocephalation” proposed by [Parker \(2000\)](#); see also [Harbitz et al., 2003](#)) for slide masses undergoing hydroplaning can explain the occurrence of isolated olistoliths and detached parts of cohesive debris flow in the form of fore- and out-runner blocks.

In the Northern Apennines, for example, brecciated substratum intervals, decimeters to meters in thickness, with soft clasts in marl layers usually occur discontinuously at the base of cohesive olistostromes over an area of about 300 km long and tens of kilometers wide ([Festa and Codegone, 2013](#); [Festa et al., 2015b, 2015c](#)). These relationships and structures suggest a non-indurated state of the underlying sediments (i.e. overridden substratum) during motion and immediately after the emplacement of olistostromes. Formation of the breccia was spatially and temporally associated with mm- to cm-thick, sub-vertical flame-shaped sedimentary injections, representing the products of fluid-overpressure driven deformation of the substratum as a result of the dynamic loading of the moving slide mass. Centimeters- to decimeters-thick “mylonite-like” shear zones ([Fig. 12C, D, E, F](#)) occur at the base of and within the olistostrome and display strong deformation fabrics in both soft- and hard-microclasts and clasts. These structures define a constriction plus flattening type (i.e., prolate plus oblate) strain ellipsoid, with a prevailing component of stretching along the direction of flow and a minor component of planar flattening due to compaction. The shear zones represent the loci of concentrated viscous deformation, which either acted in combination with basal “carpet” hydroplaning or in isolation, after dissipation of the basal fluid overpressure during the final stages of the emplacement of a slide mass (see also [Ogata et al., 2014a](#)).

Combination of hydroplaning and shearing during downslope translation of submarine landslides is an efficient mechanism for the emplacement of both basin-wide olistostromes and some “gravitational nappes” characterized by long to unusually long run-out distances achieved in relatively short time spans. This mechanism may also explain the great mobility of olistoliths up to 105–106 m<sup>3</sup> in volume, comparable in character and size to the modern out-runner blocks. Such olistoliths also occur as isolated elements in front of large olistostromes or nappe sheets (e.g., in the

External Ligurian Units), or at their base because of subsequent downslope translation of the olistostromes or nappes (e.g., Central Appalachians, see [Codegone et al., 2012a](#); Porma mélange, see [Alonso et al., 2006, 2015](#)). These kinds of olistoliths may also originate during rapid deceleration of a landslide front, whereby over-consolidated blocks with high momentum persist in their inherited motions (e.g., [De Blasio et al., 2006](#)). In modern examples, they may show tens of kilometers run-out distance, ahead of the front of submarine landslides along a less than one degree average slope (e.g., [Prior et al., 1987](#); [Nissen et al., 1999](#); [Canals et al., 2004](#); [Nielsen and Kuijpers, 2004](#)).

#### 4.2. Mesoscale kinematic indicators for mechanisms of olistostrome emplacement and translation

We describe here a wide range of mesoscale structures in discussing various deformation mechanisms of olistostromes that favor their downslope mobility. This topic is significant to better understand other factors controlling the preservation and significance of olistostromes in the evolution of orogenic belts.

Mechanisms supporting the extraordinary downslope mobility of olistostromes are explained with the application of structural geology tools in addition to classic sedimentological ones ([Ogata et al., 2016](#); [Fig. 13](#)). Mixed pure- and simple-shear mechanisms due to the coupled, cyclic action of dynamic/static loading and the differential movements of a slide mass and its internal components, produce a variety of asymmetrical structures ranging from microscopic to outcrop-scale ([Fig. 13](#)). The shape, spatial arrangement, and geometric relationships of these structures can be used as stand-alone kinematic indicators to record local differential movements between the internal slide parts or in combination for a robust interpretation of the general paleo-transport directions. These structures, specifically distributed within a slide body, are interpreted as products of soft sediment deformation developed at low confining pressures (i.e. superficial conditions) involving undrained, water-saturated, poorly- to un-consolidated sediments, both failed and eroded from the overridden seafloor. This interpretation is further supported by microscopic analyses highlighting independent particulate flow with minor or no grain breakage (see, e.g., [Ogata et al., 2014b](#)).

Close morphological similarities with ductile (and brittle–ductile) structures documented in structurally deeper metamorphic rocks allow us to adopt some of the descriptive, non-genetic terminology classically used in structural geology (e.g., [Passchier and Trouw, 2005](#)). The most common structures widely recognized in all the different examples of analyzed olistostromes include the following ([Ogata et al., 2016](#)):

- (1) Asymmetric boudinage structures ([Fig. 13\(1\)](#)) recording layerparallel extension of poorly-consolidated interbeds, having high rheological contrasts. They display characteristic pinch-and-swell geometries and a preferred alignment of phacoidal and lozenge-shaped structures. The single asymmetrical elements define the associated stress direction;
- (2) Pseudo-sigma structures ([Fig. 13\(2\)](#)), reminiscent of typical sigma structures of tectonically induced ductile deformation. These structures consist of mm- to m-sized, asymmetric, sigma shaped objects deformed by simple shear-related viscous flow. They comprise cohesive lithologies, distinctly different from the surrounding matrix, which commonly originates from the disaggregation of the transported and eroded, layered sediments;
- (3) Pseudo S–C structures ([Figs. 13\(3\)](#) and [12F](#)), comprising discrete zones (millimeters to centimeters-thick shear bands) of high shear strain (i.e. C-surfaces) that bound systematic, pervasive, sigmoidal foliation planes (i.e. S-surfaces). These planar discontinuities are commonly represented by disaggregation, deformation-and compaction-bands;
- (4) Intrafolial folds ([Fig. 13.4](#)) with their main reference surfaces traceable laterally into unfolded layers. They can be cm- to m-long and display marked thickening of their hinge zones. They show an overturned, isoclinal, sheath-type geometry in three dimensions, similar to fault-propagation folds, and may evolve into detached folds;
- (5) Low-angle shear zones ([Fig. 13\(5\)](#)), ranging in thickness from millimeters to meters and in length from centimeters to tens of meters. They can be subdivided in roughly symmetrical (with respect to the principal slip surface/interval) elements represented by damage- and core-zones, whereby transition from relatively undeformed to completely disrupted sediments (i.e. matrix) occurs in short distances. The rate and magnitude of accommodated deformation are proportional to the abundance of matrix material. Disaggregation–deformation–compaction bands are the most common features and locally occur in swarms;
- (6) Cuspitate injectites ([Fig. 13\(6\)](#)), are represented by pipe-, sheet-(i.e. linear), and wedge-shaped (i.e. planar) matrix injections. They range in thickness from millimeters to meters and in length from centimeters to tens of meters. As over-pressurized, liquefied or fluidized sedimentary matrix material penetrates into slide blocks via hydrofracturing and/or by passively flowing into the low-pressure zone, these injectites form systematically along the bounding surfaces between the matrix-rich and coherent lithologies (e.g. blocks margins, basal contact). They hence provide directional information on the shear sense and relative movements of the overpressurized matrix flow;
- (7) Duplex-type structures ([Fig. 13\(7\)](#)), are formed as a result of local imbrication of isolated blocks along flat-ramp to sigmoidal shear surfaces. They can be centimeter to tens of meters in size, and are vertically bounded above and below by major shear surfaces merging laterally into a single shear plane;
- (8) Rootless folds ([Fig. 13\(8\)](#)), representing completely detached, isolated slump-type fold hinges, dispersed within a fine-grained matrix. In three dimensions they resemble asymmetric, isoclinal sheath folds with marked thickening of their hinge zones. They commonly include sedimentary matrix injections along their outer-arcs, limbs and cores. These structures, along with the intrafolial counterparts (see above), are the most effectively

used kinematic indicators for interpreting paleo-transport directions (e.g. [Lucente and Pini, 2003](#); [Strachan and Alsop, 2006](#); [Ogata et al., 2014b](#)).

In summary, the identification of these structures in outcrops helps us correctly interpret the general and local slide kinematics of olistostromes, and hence complementing the small-scale datasets provided by drill cores, borehole logging, and geophysical imaging.

## 5. Discussion: olistostromes as markers of tectonic events

Olistostromes and mass-transport deposits are predominantly gravitational in origin and occur widely in ancient orogenic belts and exhumed subduction–accretion complexes. Close relationships between different types of olistostromes and tectonic settings of their formation ([Camerlenghi and Pini, 2009](#); [Festa et al., 2010a, 2012b](#)), the degree of their preservation in the geological record, and the analysis of their internal fabric revealing the potential mechanisms of their emplacement allow us to evaluate the significance of olistostromal occurrences in the evolution of ancient orogenic belts and exhumed subduction complexes. There remain two major questions: whether or not olistostromes represent the local markers of regional tectonic events, and whether their emplacement is a recurrent event in a particular tectonic setting.

The discontinuous distribution of olistostromes in space and time within ancient orogenic belts and subduction–accretion complexes may reflect significant physical and mechanical changes in their depositional settings of formation. Tectonic events commonly show a strong correspondence with transgressive–regressive sedimentation cycles, which are, in turn, in agreement with the local and eustatic sea level variations ([Fig. 14](#); see e.g., [Vail and Mitchum, 1980](#); [Haq et al., 1987](#); [Haq and Schutter, 2008](#); [Snedden and Liu, 2010](#); see [Fig. 14](#)). Eustatic sea level fluctuations commonly result in the formation of global unconformities at sequence boundaries of sedimentary cycles. Thus, except for local cases, the length of each time interval of olistostrome emplacement corresponds to a single tectono-sedimentary cycle ([Fig. 14](#)). Such a cycle can be defined, for example in active margins, by a time window between the nappe emplacement and the related accumulation of regressive sedimentary sequences and the subsidence initiation that facilitates mass transport processes and the formation of olistostromes in frontal parts of advancing nappe sheets (e.g., [Scherba, 1989](#); [Lucente and Pini, 2008](#); [Festa et al., 2015c](#)). The main stages of olistostrome emplacement in the Circum-Mediterranean Region (i.e., Apennines, Betic Cordillera, Rif, and Hellenides–Albanides) and the Alps (the Ligurian Alps, Carpathian, Pyrenees, and Caucasus) during the late Cenozoic may be attributed to the same or closely similar time intervals ([Fig. 14](#)), which correspond to main tectonic stages and regional-scale transgressive–regressive cycles ([Scherba, 1989](#)). Although olistostromes developed during each single tectonic stage do not occur everywhere in all these regions, the majority of them are confined to ([Fig. 14](#)) the middle Eocene, late Eocene–early Oligocene, late Oligocene–early Miocene, late–early Miocene (Burdigalian), late Miocene (late Messinian)–early Pliocene (see, e.g., [Scherba, 1989](#); [Pini, 1999](#); [Lucente and Pini, 2003, 2008](#); [Camerlenghi and Pini, 2009](#); [Festa et al., 2010b](#); [Ślaczka et al., 2012](#); [Ogata et al., 2012b](#); [Festa and Codegone, 2013](#); [Festa et al., 2013](#); [Ogata et al., 2014a](#); [Festa et al., 2015b, 2015c](#)). We thus posit that these specific time windows correspond to different stages of intra-continental tectonics, starting with the Eocene collision between Adria and Europe, after the closure of the Tethyan oceanic realm. Although, in general, some structural differences exist between the northern and southern branches of the Circum-Mediterranean region, each stage corresponds to a main phase of nappe transport toward the foreland until the final closure of Neothetys (e.g., [Scherba, 1989](#); [Lucente and Pini, 2008](#); [Camerlenghi and Pini, 2009](#); [Festa et al., 2010b](#) and reference therein). Similar observations can be made in different orogenic belts, and within the remnants of exhumed ancient subduction–accretion complexes in the Circum-Mediterranean Region, Appalachians and Circum-Pacific Region, independently of the ages of different tectonic events and the characteristics of different orogenic stages ([Fig. 14](#)).

In all these orogenic belts, the temporal and spatial distribution of different types of olistostromes commonly represent different tectonic stages within the Wilson cycle evolution of ocean basins, from the early rift–drift to later subduction, collision, and orogenic exhumation ([Fig. 14](#); [Dilek, 2003](#); [Dilek and Newcomb, 2003](#)). One of the best examples of such time-progressive olistostrome development can be given from the Circum-Mediterranean region, where different types of olistostromes reflect different stages of the tectonic evolution of the Alpine Tethys. For example, different phases of olistostrome emplacement have been well documented to be associated with the Late Jurassic rifting stage during the opening of the Alpine Tethys (e.g., [Elter and Trevisan, 1973](#); [De Libero, 1998](#); [Cieszkowski et al., 2012](#); [Clerc et al., 2012](#); [Balestro et al., 2015a](#); [Festa et al., 2015a](#); [Tartarotti et al., 2015](#)), the Late Jurassic–Early Cretaceous collapse of passive margin carbonate platforms (e.g., [Castellarin, 1972](#); [Cecca et al., 1981](#), [Fazzuoli et al., 1985](#); [Bernoulli, 2001](#); [Graziano, 2001](#); [Ślaczka et al., 2012](#)), the Late Cretaceous–early Paleogene subduction stages (e.g., [Elter and Trevisan, 1973](#); [Marroni and Pandolfi, 2001](#)), and the Eocene–Miocene collisional and intra-collisional stages (e.g., [Abbate et al., 1970](#); [Bettelli et al., 1989](#); [Pini, 1999](#); [Panini et al., 2002](#); [Lucente and Pini, 2003, 2008](#); [Camerlenghi and Pini, 2009](#); [Festa et al., 2010b](#); [Remitti et al., 2011](#); [Codegone et al., 2012b](#); [Ogata et al., 2012c, 2014a](#); [Festa et al., 2015c](#)).

In the Central Appalachians, different types of olistostromes and olistoliths mark a continuum of tectonic events ([Fig. 14](#)) from the early stages of subduction of the Laurentian continental margin beneath an intraoceanic island arc system in the middle Ordovician (e.g., [Epstein et al., 1972](#); [Lash and Drake, 1984](#); [Lash, 1987](#); [Ganis et al., 2001](#); [Ganis and Wise, 2008](#); [Codegone et al., 2012a](#)) to the early–late Ordovician and late Ordovician precursory olistostrome



emplacement (Martinsburg Formation) in a foreland basin of the initial Appalachian orogenic belt during the collision stage (e.g., [Root and MacLachlan, 1978](#); [Lash and Drake, 1984](#); [Lash, 1987](#); [Codegone et al., 2012a](#)). A comparable

olistostromal sequence characterizes the Taconic foreland basin extending from the New York State to Vermont and Southern Quebec ([Berry, 1962](#); [Bird, 1963](#); [Stevens, 1968](#); [St-Julien and Hubert, 1975](#); [Lash, 1985](#); [Vollmer, 1980](#); [Bosworth and Vollmer, 1981](#); [Bosworth, 1989](#); [Schroetter et al., 2006](#); [Festa et al., 2012a](#)), where they record different stages of the Taconic orogeny. On the East Coast of New Zealand in the SW Pacific, [Delteil et al. \(2006\)](#) have documented the occurrence of successive tectonic events marked by different types of olistostromes spanning a time interval from the Eocene to the early Miocene.

The recurrence time of olistostrome emplacement thus appears to be directly related to the frequency of tectonic stages, their magnitude, and their control on depositional cycles ([Fig. 14](#)). In the continuum of deformation characterizing different stages of the orogenic belt evolution, the seemingly discontinuous distribution of olistostromes with different ages in different tectonic settings represents, on the contrary, a genetic phylum (see also [Camerlenghi and Pini, 2009](#)) of direct relationships between tectonic events as primary triggering mechanisms and olistostromes. Hence, olistostromes can be interpreted as fundamental markers of tectonic events ([Fig. 14](#)), and their study can provide a powerful tool for basin analysis at various scales.

The recurrence interval of olistostromes in the geological record may be unrevealed with varying degrees of accuracy, based on the well constrained timing of olistostrome emplacement, subsequent tectonic deformation events, and metamorphic history of the depositional environment of olistostromes. Although the emplacement of an each single gravitational body is an instantaneous and episodic event, the recurrence time for olistostrome and olistolith emplacement may vary from thousands of years to several millions of years (e.g., [Scherba, 1989](#); [Graziano, 2001](#); [Delteil et al., 2006](#); [Ganis and Wise, 2008](#); [Camerlenghi and Pini, 2009](#); [Festa et al., 2015c](#)). This time-scale may be strongly affected by the duration of each regional-scale, single tectonic event, or more specifically by the combination of eustatic and geodynamic processes (e.g., sedimentary transgressive–regressive cycles, global climate changes, gas-hydrates dissociations). Olistostromes are, therefore, 3-D tape-recorders for time-progressive evolution of orogenic belts and subduction–accretion systems.

## 6. Concluding remarks

In this comparative synthesis we have examined the occurrence and the internal structure of different types of olistostromes and olistostromal deposits in various orogenic belts (with a focus on the Circum-Mediterranean, Circum-Pacific and Appalachian regions), and have addressed some of the long-lasting “olistostrome debates”. This study shows that different types of olistostrome and their characteristic features are closely related to the specific tectonic settings of their formation. The systematic documentation of their blocks-in-matrix fabric, the nature and compositions of their blocks, and the lithologies, compositions and grain sizes of their matrix provide critical information on different types of deformation processes that these chaotic deposits experienced during and after their emplacement. In particular, their lithological makeup presents key evidence for determining the internal architecture of exhumed subduction–accretion complexes. Rearrangement of sediment delivery mechanisms and the types and volumes of material entering into accretionary wedges and subduction channels may vary significantly during discrete evolutionary phases and in specific parts of an accretionary wedge.

Olistostromes commonly represent complex gravitational bodies, consisting of different parts; they do not exclusively represent the products of a single depositional event. Their internal structure reflects a close superposition in space and time of different gravitational bodies, forming mass-transport complexes that are comparable in size with those observed in modern submarine settings. However, in contrast to some of the modern examples in which submarine landslides are mainly associated with gravitational instability along passive margins, ancient olistostromes are mainly preserved in the geological record of active margin settings. This phenomenon is largely due to their nature and composition, to the physiography of the depositional setting, and to downslope transformation mechanisms related to long run-out distances. Slide masses may reach extraordinary distances of translation from the source area (tens to hundreds of kilometers) that may not be evident from their internal fabric and compositions. This apparent discrepancy may stem from the combination of hydroplaning and undrained simple-pure shearing mechanisms, which occur during downslope translation at the base of and inside the gravitational body, respectively. Detailed characterization of a chaotic deposition through the use of sedimentological and structural nomenclature and terminology may be exceedingly hard to do in modern mass-transport deposits due to the limited resolution of the geophysical methods. Thus, a multiscale structural analysis, supported by sedimentological and stratigraphic observations, carried on continuous three-dimensional on-land outcrops, is commonly needed to document thoroughly the effects of strain partitioning caused by differential movements within a slide mass ([Ogata et al., 2016](#)).

Olistostromes are critical geological bodies to examine carefully in studying ancient orogenic belts and exhumed subduction–accretion complexes with a history of multi-stage deformational events. They provide excellent markers for tectonic and climatic events, and their ages and structures can be used effectively for basin analysis and modeling. Careful documentation and understanding of the overall architecture of olistostromes, their meso- to map-scale internal structures, and the mechanisms of their downslope deformation and emplacement are highly important and relevant for better understanding of modern submarine landslides and potential geohazards associated with them.



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Andrea Festa is an Associate Professor at the Department of Earth Sciences at the University of Torino (Italy). He received his PhD in Earth Sciences from the University of Genova, Italy in 2000. After research appointments (University of Catania, University of Torino — ENI-E&P), and a Researcher position at National Research Council of Italy (CNR-IGG), he became Researcher in 2005 at the University of Torino and then Associate Professor. He has been Visiting Professor at Miami University (OH-USA). His major research interests are the relationships between tectonics and sedimentation, the processes of mélanges formation in different geodynamic settings in the Circum-Mediterranean Region, Appalachians, US-Western Cordillera, and Argentinean Andes, and the reconstruction of intra-oceanic tectono-stratigraphic setting in exhumed ophiolitic belts. Geological mapping represents the base and the most important approach for his research activity.



Kei Ogata is a research fellow at the Department of Physics and Earth Sciences of the University of Parma (Italy). He received his PhD in Earth Sciences at the same Department in 2010. After postdoctoral appointment at the University Centre in Svalbard, Norway, he moved back to Parma in 2013. He is now Assistant Professor at the VU University, Amsterdam (The Netherlands). His research interests focus on the understanding of the interplay between sedimentary and tectonic deformation mechanisms in sedimentary rocks at the various scales, from the early syn-depositional processes to later post-lithification fracturing.



Gian Andrea Pini is an Associate Professor at the Dipartimento di Matematica e Geoscienze at the Università di Trieste (Italy). He received his PhD in Geology from the Università di Bologna and Modena Consortium in 1987. After postdoctoral appointments, he became Permanent Researcher and, then, Associate Professor at the Università di Bologna until 2013, when he moved to the Università di Trieste. His major research interests are in the processes originating mélanges, mud diapirs and mass transport deposits in the Circum-Mediterranean Chains, New Zealand, and Argentine Precordillera.



Professor Yildirim Dilek is a structural geologist–geochemist working on the structure, petrology and tectonic evolution of modern and ancient oceanic lithosphere, mélanges and subduction-accretionary complexes, collisional–extensional granitoids, and rifted continental margins. He received his PhD from the University of California. He has served as the Vice President of the International Union of Geosciences (IUGS) since 2012, and is currently serving as the Chair of the Mineralogy-Geochemistry-Petrology-Volcanology Division of the Geological Society of America. He has been a Visiting Professor at Ecole Normale Supérieure in Paris (France), University of Kanazawa (Japan), China University of Geosciences—Beijing (China), University of Tohoku in Sendai (Japan), and the Chinese Academy of Geological Sciences in Beijing (China) during the past ten years. He has worked extensively in the Alpine, Peri-Mediterranean, Tibetan–Himalayan, Qinling, North American Cordillera, and Norwegian Caledonide orogenic belts. His most current research project topics include melt evolution of ophiolitic peridotites, diamonds and recycled mantle, and extensional tectonics and magmatism in young orogenic belts. He is a co-leader of Project IGCP-649, Diamonds and Crustal Minerals in Ophiolitic Peridotites: Global Models for Mantle Dynamics. He has published over 200 papers and 24 edited books and special issues of journals. His other interests include geopolitics, history and landscape of the Silk Road, jazz and soccer.



Juan Luis Alonso is a Reader at the Department of Geology at the University of Oviedo (Spain), where he received his PhD in geology in 1985. His research interests include fault-related folding, inherited structures (fault and fold reactivation) and slumping processes. He has worked in the Variscan and Alpine deformation in Iberia and in the Andean and pre-Andean structure in the Argentinean Andes.



Products	Sub-products	Nature of blocks	Mechanisms	Lithological unit involved	Contacts with host rocks	Processes
Olistostrome	Allolistostrome Sedimentary mélange	Exotic and native	Mixing	Sedimentary Metamorphic Igneous	Stratigraphic	Sedimentary (Mass transport)
	Endolistostrome	Native (Intra-formational)	Stratal disruption	Sedimentary		

Table 1

Characteristic features of the two end members of olistostromes, based on the nature of blocks and mechanism of formation.

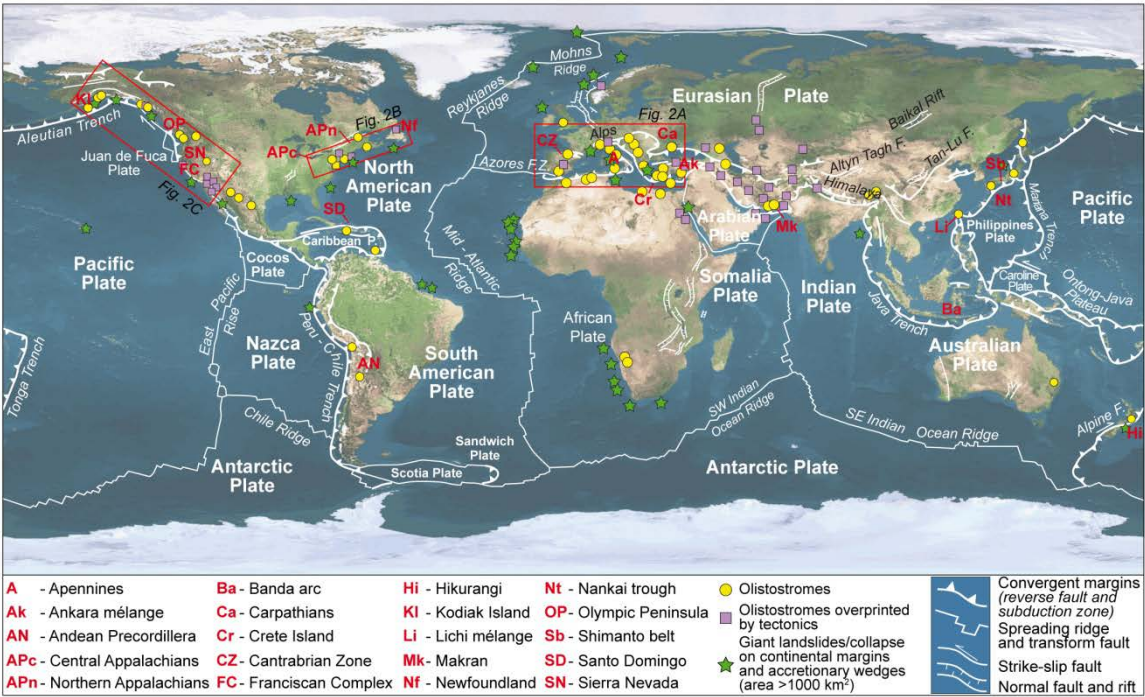


Fig. 1. Global distribution of main modern (from Mienert et al., 2003) and ancient (from Raymond, 1984; Camerlenghi and Pini, 2009; Festa et al., 2010a) olistostromes and submarine landslide deposits (modified after Dilek et al., 2012).

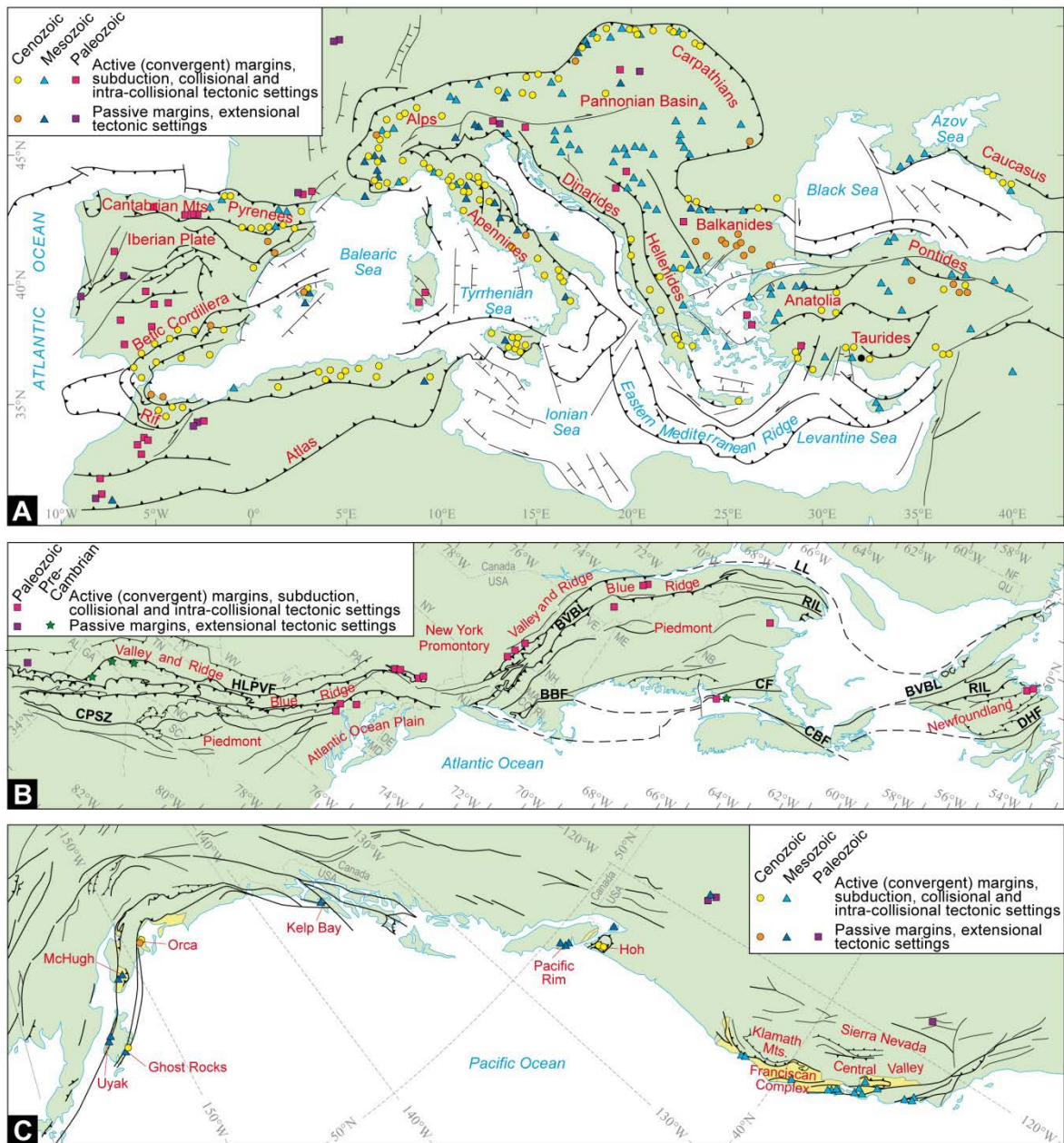


Fig. 2. Distribution of olistostromes and MTDs in the (A) Circum-Mediterranean Region (modified from Camerlenghi and Pini, 2009), (B) Appalachian orogenic belt (modified from Rast and Horton, 1989; Festa et al., 2010a; Codegone et al., 2012a), and (C) Western coast of the USA and Canada (from Raymond and Terranova, 1984; Cowan, 1985; Brandon, 1989; Blome and Nestell, 1991; Aalto, 2014; Raymond and Bero, 2015; Wakabayashi, 2015), distinguished by the tectonic setting of their formation and emplacement age. Tectonic lineaments in (A) from Camerlenghi and Pini (2009), Vezzani et al. (2010) and Quintana et al. (2015), in (B) from Hibbard et al. (2006) and Hatcher et al. (2007), in (C) from Wheeler and McFeely (1991) and Reed et al. (2005). BBF: Bloody Bluff Fault, BVBL: Baie Verte–Brompton Line, CBF: Chedabucto Fault, CF: Caledonia Fault, CPSZ: Central Piedmont Shear Zone, DHF: Dover–Hermitage Bay Fault, HLPVF: Hollins Line–Pleasant Valley Fault System, LL: Logane Line, RIL: Red Indian Line.

Types of Olistostrome related to:	Geodynamic environment	Processes	Triggering mechanisms	Products	Mesoscale characteristics
<b>Passive margin</b>	Passive margins (during and after rifting)	MTP (debris avalanches and flows, etc.)	Tectonic	MTD (megabreccias, breccias, olistolith fields, debrites, slide blocks, etc.)	Chaotic angular clasts (cm to >10 m) in fine-grained (pelitic) matrix
<i>Collapse of platform margins</i>					
<i>Mass-transport deposits at the ocean-continent transition (OCT)</i>	Ocean-continent transition	SSD and MTP with related progressive deformation from slumping to debris flows, to gravitational sliding	Tectonic, sedimentary	MTD, olistostromes with continent rock olistoliths (tens of meters to several km slide blocks) in a matrix of oceanic origin	Chaotic monomictic to polymictic brecciated (matrix-supported) masses (including native, extra-basinal and/or exotic blocks)
<i>Intra-oceanic downslope mass-transport deposits</i>	Oceanic realm (mid-oceanic ridge, seamounts)	MTP (debris avalanches and flows, etc.)	Tectonic	MTD (breccias, olistolith fields, debrites, slide blocks, etc.)	Chaotic angular clasts (cm to >5 m) in fine-grained pelagic matrix
<b>Convergent margins and oceanic crust subduction</b>	Subduction (at the front of the wedge) and fore-arc basins	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, BrFm, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (including native, extra-basinal and/or exotic blocks)
<i>Mass-transport deposits at the wedge and retro-wedge front</i>	Obduction settings (from intra-oceanic to marginal stage)			MTD, BrFm, olistostromes (olistoliths, olistolith fields, slide blocks, breccias, debrites)	
<i>Obduction-related olistostromes</i>					
<b>Collisional and intracontinental deform.</b>					
<i>Mass-transport deposits at the wedge front</i>	At the base or at the front of intra-continental thrust sheets or nappes	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (from matrix-supported cm-to m in size blocks to clast supported >10 m blocks and olistoliths)
Precursory olistostromes		TSD: fault-to fold-related, fluidification (overprinting previous mass-wasting-related deformation)	Tectonic, sedimentary	Mélanges (exotic blocks being commonly recycled from other previously formed olistostromes)	Chaotic BIM fabric overprinted by tectonic deformation and shearing
Olistostromal carpet					
<i>Intra-nappe</i>					
Sedimentary	Within intra-continental thrust sheets or nappes	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (blocks of intra-basinal origin)
Tectono-sedimentary		TSD: fault-to fold-related, fluidification (overprinting previous mass-wasting-related deformation)	Tectonic	BrFm; mélanges (exotic blocks being commonly recycled from other previously formed olistostromes)	Structurally ordered BIM fabric (parallel orientation of blocks and matrix features – i.e. pseudo-bedding)
<i>Epi-nappe</i>					
Sedimentary	A top of intra-continental thrust sheets or nappes (e.g. piggy back, top thrust basins)	MTP (debris flows and avalanches, slumps, slides, etc.)	Tectonic, sedimentary	MTD, olistostromes (olistoliths, olistolith fields and swarm, slide blocks)	Chaotic BIM fabric (originated from the succession tectonically imbricated in the thrust-sheet)
Tectono-sedimentary		TSD (overprinting previous mass-wasting-related deformation)	Tectonic, sedimentary	BrFm; mélanges (exotic blocks being commonly recycled from other previously formed olistostromes)	Structurally ordered BIM fabric
BIM – Block-in-matrix BrFm – Sedimentary Broken Formation		MTD – Mass-transport deposits MTP – Mass-transport processes		SSD – Soft sediment deformation TSD – Tectonic stratal disruption	

Table 2  
Subdivision and classification of olistostromes (modified after [Festa et al., 2010a, 2012b](#)).



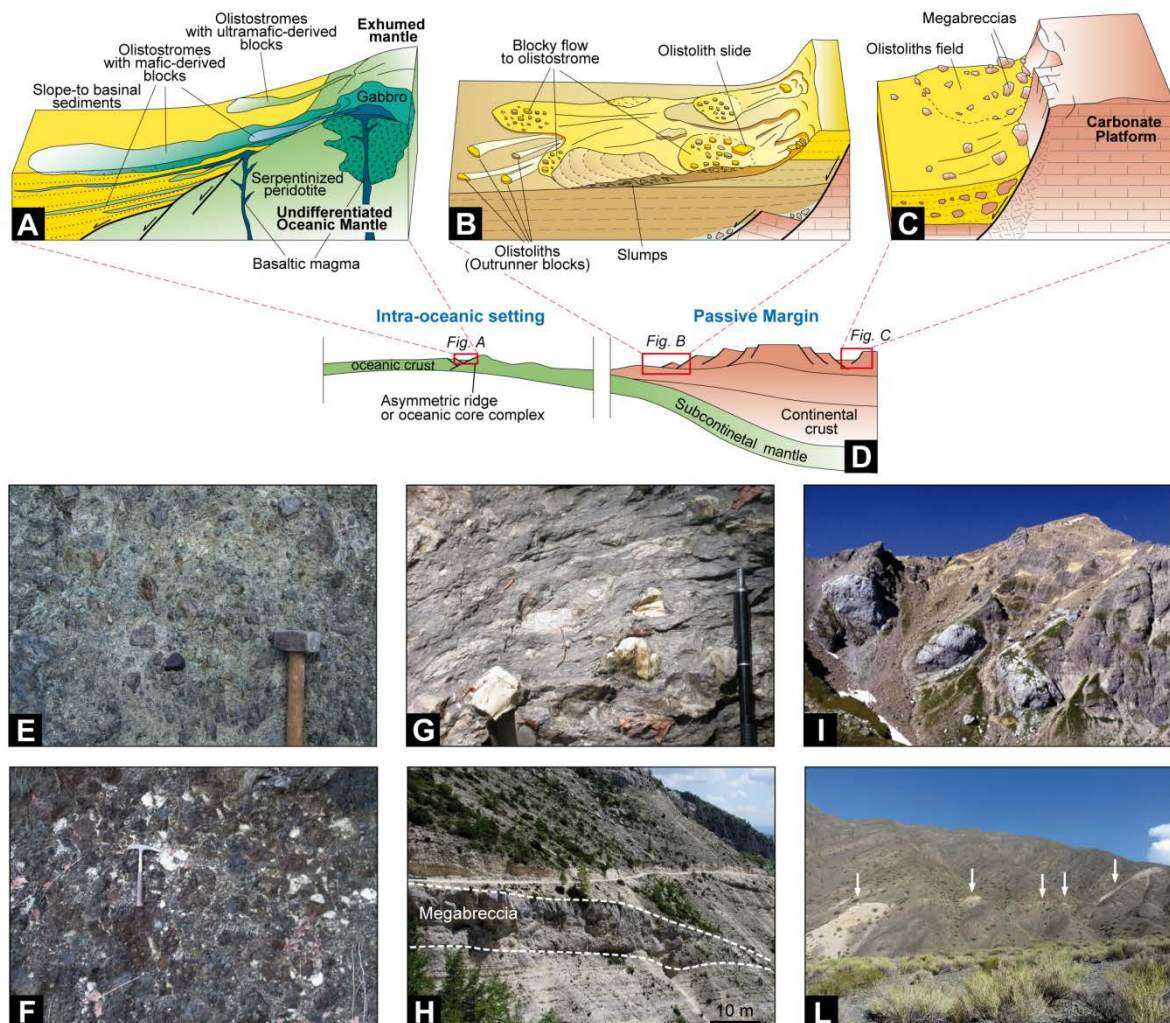


Fig. 3. Conceptual models for the formation and emplacement mechanisms of olistostromes and ancient MTDs associated with various stages of rift–drift and passive margin evolution: (A) Intra-oceanic settings, close to sea-floor spreading ridge (modified from [Balestro et al., 2015b](#), [Festa et al., 2015a](#)); (B) passivemargin at ocean–continent transition (OCT) (modified from [Prior et al., 1984](#)), and (C) collapse of a carbonate platform (modified from [Bosellini, 1998](#)). (D) Olistostrome sites depicted in the three conceptual models in A through C (from [Balestro et al., 2015b](#)). Field examples showing: (E) rounded to angular clasts of ultramafic rocks in a fine- to medium grained matrix of the same composition, derived from downslope mass-transport processes in intra-oceanic settings (Ligurian Units, Northern Apennines, Italy); (F) clast-supported debris flow consisting of polymictic rounded to angular clasts of ultrabasic rocks and an oceanic cover succession (Ligurian Units, Northern Apennines, Italy; courtesy of E. Barbero); (G) muddy debris flow body (olistostrome) of the Modino basal complex (Northern Apennines, Italy; courtesy of C.C. Lucente), showing the flow-related deformational features of non-consolidated carbonate clasts in an argillaceous matrix; (H) tens of meters thick, channelized megabreccias, composed of platform blocks within a white-colored, hemipelagic Upper Cretaceous calcilutite (Majella Mountain, Central Apennines, Italy; see [Festa et al., 2014a](#)); (I) Upper Cretaceous megabreccia of calcareous limestone (Muttetkopf, Calcareous Alps, Austria; see [Ortner, 2001](#); [Amerman et al., 2009](#)). The mountainside is about 300 m high; (L) blocks (tens of meters in size) of white-colored platform carbonate (white arrows) in a dark, Ordovician shale matrix (Los Ratonés Valley, Argentina Precordillera). The mountainside is about 1000 m high.



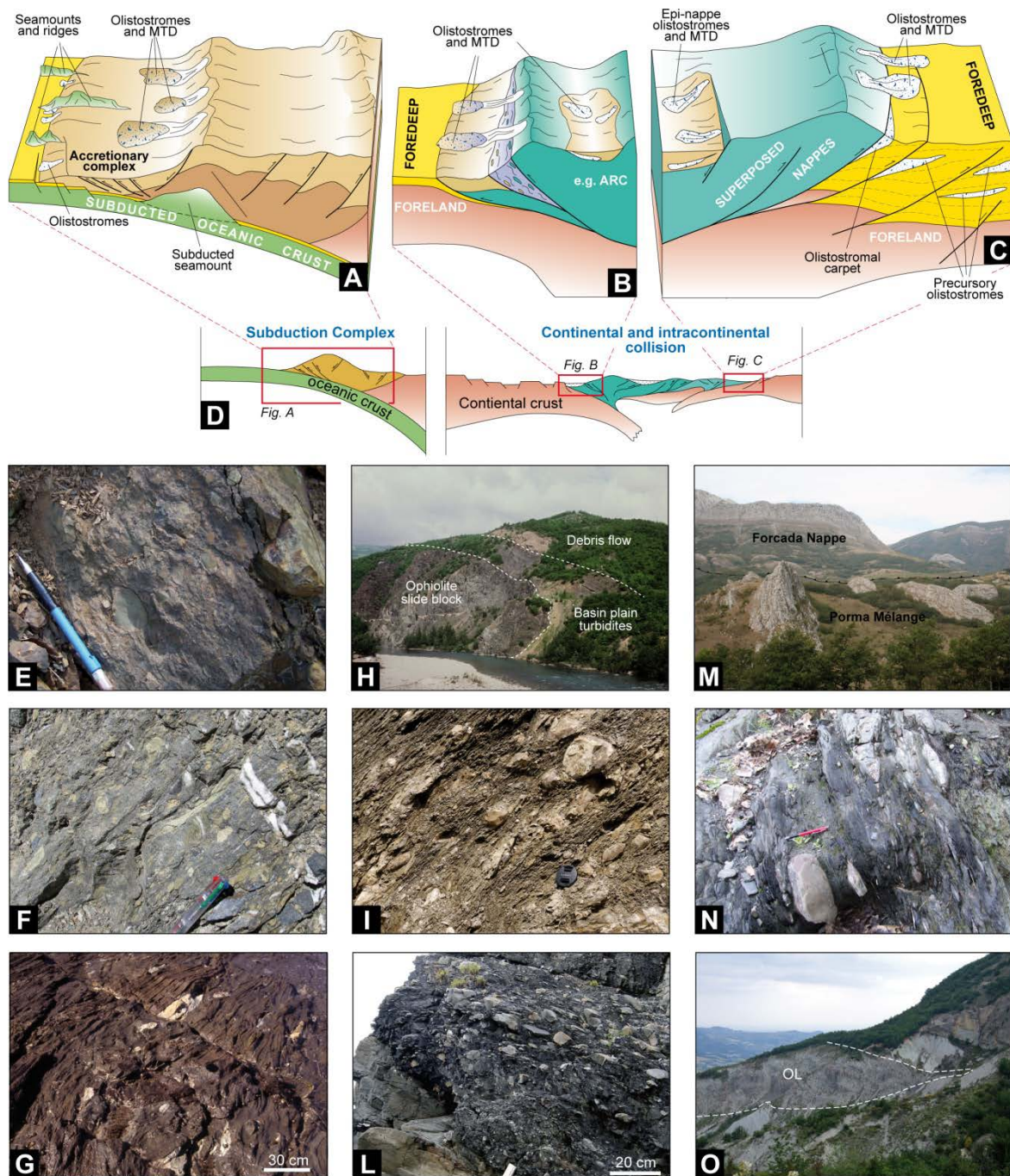


Fig. 4. Conceptual model for the formation and emplacement of olistostromes and ancient MTDs associated with (A) convergent margins and oceanic crust subduction (modified from von Huene et al., 1989; Festa et al., 2012b) and (B) collisional and (C) intra-collisional tectonic settings (modified from Festa et al., 2012b, 2015c). (D) Olistostrome sites depicted in the three conceptual models in A through C (from Marroni et al., 2010; Festa et al., 2013) and accretion of a lithospheric mantle (modified from Bortolotti et al., 2013). Field examples of olistostromes associated with convergent margin tectonics and oceanic crust subduction: (E) rounded to elongated blocks (deep-water limestone, silicified mudstone, and siltstone) included in a brecciated matrix (Berks County, Central Appalachians, Pennsylvania) formed by debris flow in front of the middle Ordovician accretionary wedge (see Lash, 1987; Codegone et al., 2012a); (F) close-up of clast-supported breccia (Panoche Road, Franciscan Complex, California), with clasts of metavolcanic and metagraywacke rocks, created by debris flows shed into a trench (see Wakabayashi, 2012); (G) coastal exposure of the Pahaoa olistostrome (Glendu Rocks, East Coast of New Zealand); (H) overturned succession showing a large ophiolitic block (several hundreds of meters wide) embedded in a turbiditic sequence, (Upper Cretaceous Casanova Complex, Northern Apennines, Italy; courtesy of E. Mutti); (I) debris flow deposit with blocks of an oceanic cover succession in a sheared, shaly matrix (Casanova Complex, Northern Apennines, Italy); it formed at the frontal toe of the Ligurian Accretionary Wedge (i.e., Alpine retro-wedge) in the Late Cretaceous. Field examples of olistostromes associated with

collisional and intra-collisional tectonics: (L) Precursory olistostrome of the uppermost portion of the Oligocene Macigno Costiero Formation (Cinque Terre area, La Spezia, Northern Apennines, Italy); (M) olistostromal carpet in front of the Bodon and Forcada Nappes showing large olistoliths inside the Carboniferous Pormamélange (Cantabrian Zone, North Spain; see [Alonso et al., 2006, 2015](#)); (N) olistostromal carpet displaying a debris flowdeposit emplaced in the footwall of the Taconic Allochthon and showing fabric elements of tectonic overprint. Exotic blocks showlenticular shapes produced by tectonic shearing (Hoosic River at Schaghticoke Gorge, eastern NY, Central Appalachians — USA); (O) panoramic view of an epi-nappe olistostrome, emplaced within the late Oligocene–early Miocene wedge-top basin succession atop of the Ligurian Units (Northern Apennines, Italy; see [Festa et al., 2015b, 2015c](#)).

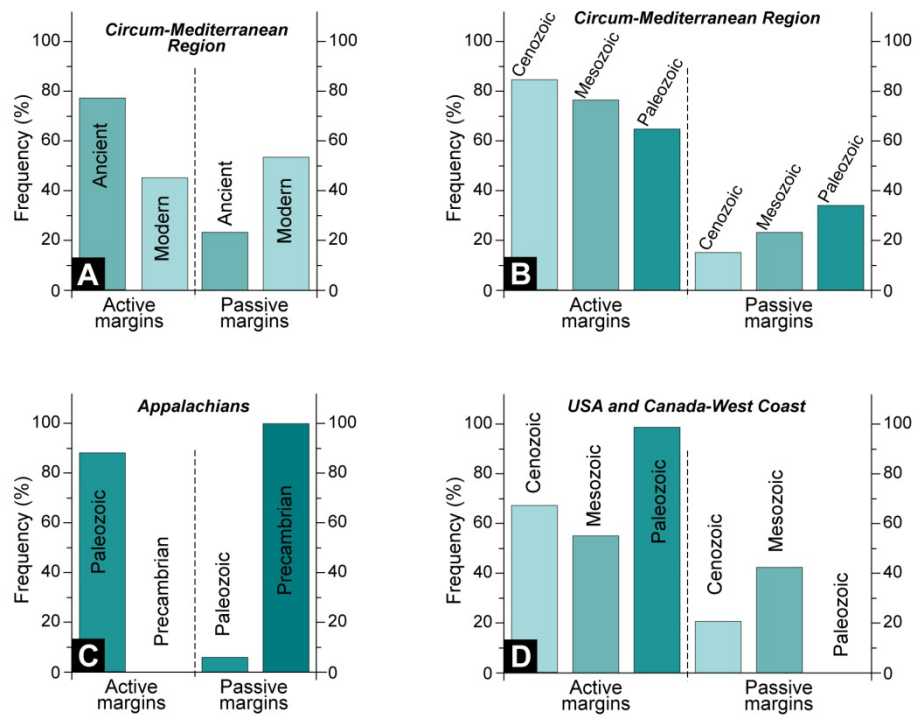


Fig. 5. (A) Comparative diagrams for the frequency of preservation of ancient olistostromes/MTDs and modern submarine landslides in passive and active (convergent) margins in the Circum-Mediterranean Region (data of modern examples from [Urgeles and Camerlenghi, 2013](#)). (B, C, D) Comparative diagrams for the frequency of preservation of olistostromes in passive and active (convergent) margins through time in the Circum-Mediterranean Region (B), Appalachians (C) and the West Coast of the USA and Canada (D).



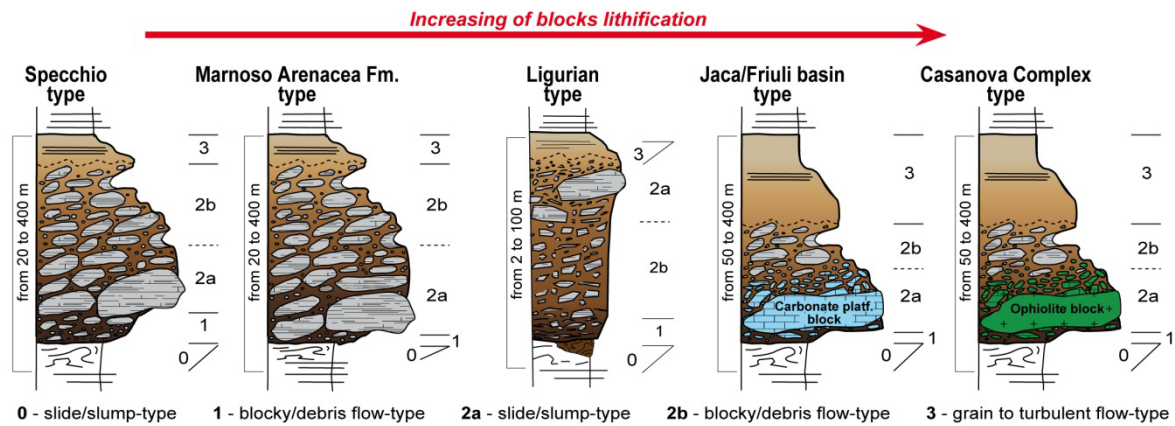


Fig. 6. Conceptual stratigraphic logs summarizing different types of studied olistostromes and MTDs. The recurrent structural facies associations are numbered from the base to the top. These schematic representations refer to inferred depocenters, in which the entire vertical succession is likely preserved.

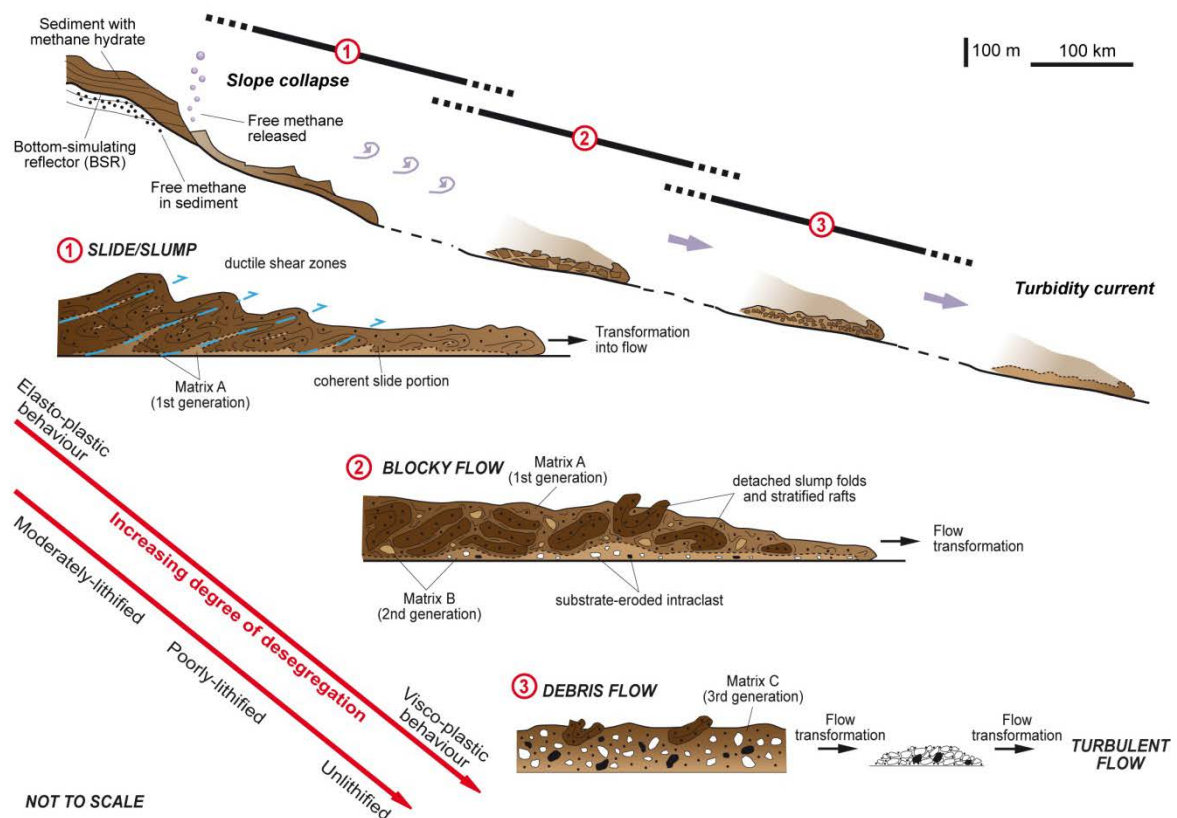


Fig. 7. Dynamic classification of mass transport deposits showing genetic and evolutionary relationships among various processes responsible for their formation (modified from Nisbet and Piper, 1998; Ogata et al., 2012b). Slide/slump and turbidity currents represent the end-members of such broad spectrum of geological processes. Blocky-flow deposits (Mutti et al., 2006), which are similar to debris-flow deposits, except for carrying over-sized slide blocks, represent a transitional type between slumps and debris-flow deposits.

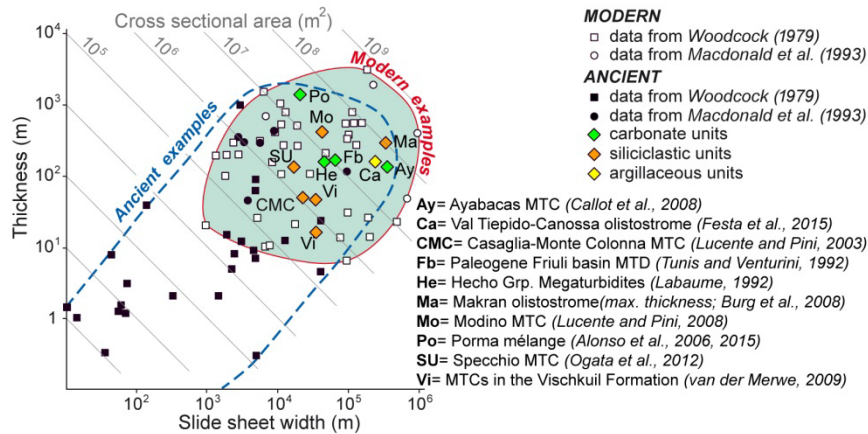


Fig. 8. Logarithmic diagram of average thickness versus average width of submarine landslides, showing relative sizes of ancient (dashed blue line envelope) and modern (solid red line envelope) olistostromes, MTDs and MTCs. Modified from Woodcock (1979), Macdonald et al. (1993), Lucente and Pini (2003), Ogata et al. (2014a).

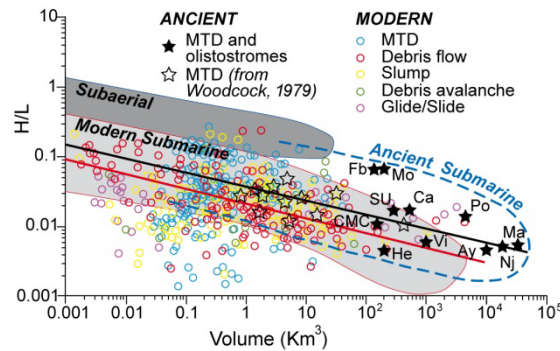


Fig. 9. Logarithmic diagram of Height of fall to horizontal runout (H/L) ratio versus volume for modern examples of subaerial and submarine landslide types (modified from Hampton et al., 1996; De Blasio et al., 2006; Urgeles and Camerlenghi, 2013; Calvès et al., 2015) and ancient submarine MTDs and olistostromes (based on extrapolation of field observations). Black and red lines show a power fit to ancient MTDs/olistostromes and debris flows data, respectively. Key to lettering: Ay: Upper Cretaceous Ajabacas MTC (Callot et al., 2008), Ca: Oligocene–Miocene Val Tiepido Canossa olistostrome (e.g., Festa et al., 2015c), CMC: Miocene Casaglia–Monte della Colonna MTC (Lucente and Pini, 2003), Fb: Paleocene Friuli Basin MTD Units (Tunis and Venturini, 1992; Ogata et al., 2014b), He: Eocene Hecho Group megaturbidites (Labaume, 1992; Ogata et al., 2012b), Ma: Miocene Makran olistostrome (Burg et al., 2008), Mo: Miocene Modino MTC (Lucente and Pini, 2008), Nj: Nataraja (Calvès et al., 2015), Po: Carboniferous Porma mélange (Alonso et al., 2006, 2015), SU: Oligocene Specchio Unit MTC (Ogata et al., 2012c), Vi: Permian MTCs in the Vischkuil Fm. (van der Merwe et al., 2009).

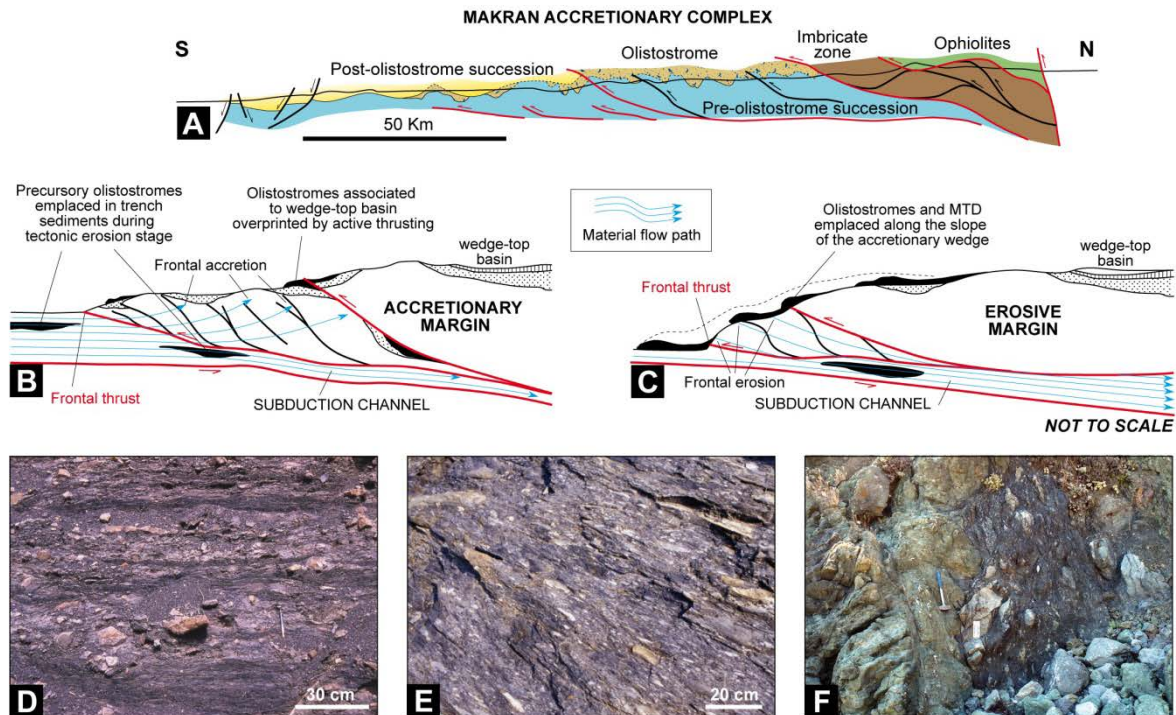


Fig. 10. (A) Simplified geological cross-section of the Makran accretionary complex (modified from Burg et al., 2008), showing the emplacement of a basin-scale olistostrome along the slope of the frontal wedge. Note that part of the olistostromal body is involved in post-depositional thrusting. (B–C) Conceptual cross-sections of accretionary (B) and erosive (C) margins (modified from Festa et al., 2015c) showing different paths of material flow (from Vannucchi et al., 2012) and inferred emplacement trajectories of old and contemporaneous olistostromes and MTDs along the slope of an accretionary wedge. (D) Upper Cretaceous olistostrome in the external Ligurian Units, flattened and slightly deformed by compaction and horizontal shortening and reverse faulting (Northern Apennines, Italy); (E) strongly flattened clasts of debris flow deposits within an olistostromal carpet at the base of the Ligurian nappe and overlying the Macigno foredeep complex (Northern Apennines, Italy); (F) large block of sheared basic metavolcanic rock ("greenstone" beneath hammer) and smaller phacoids of sandstone (e.g., beneath scale) in a sheared sandstone and mudrock matrix (Heaven's Beach mélange, Franciscan Complex, California; see Raymond and Bero, 2015). Hammer handle ~32 cm in length as a scale (courtesy of L.A. Raymond).



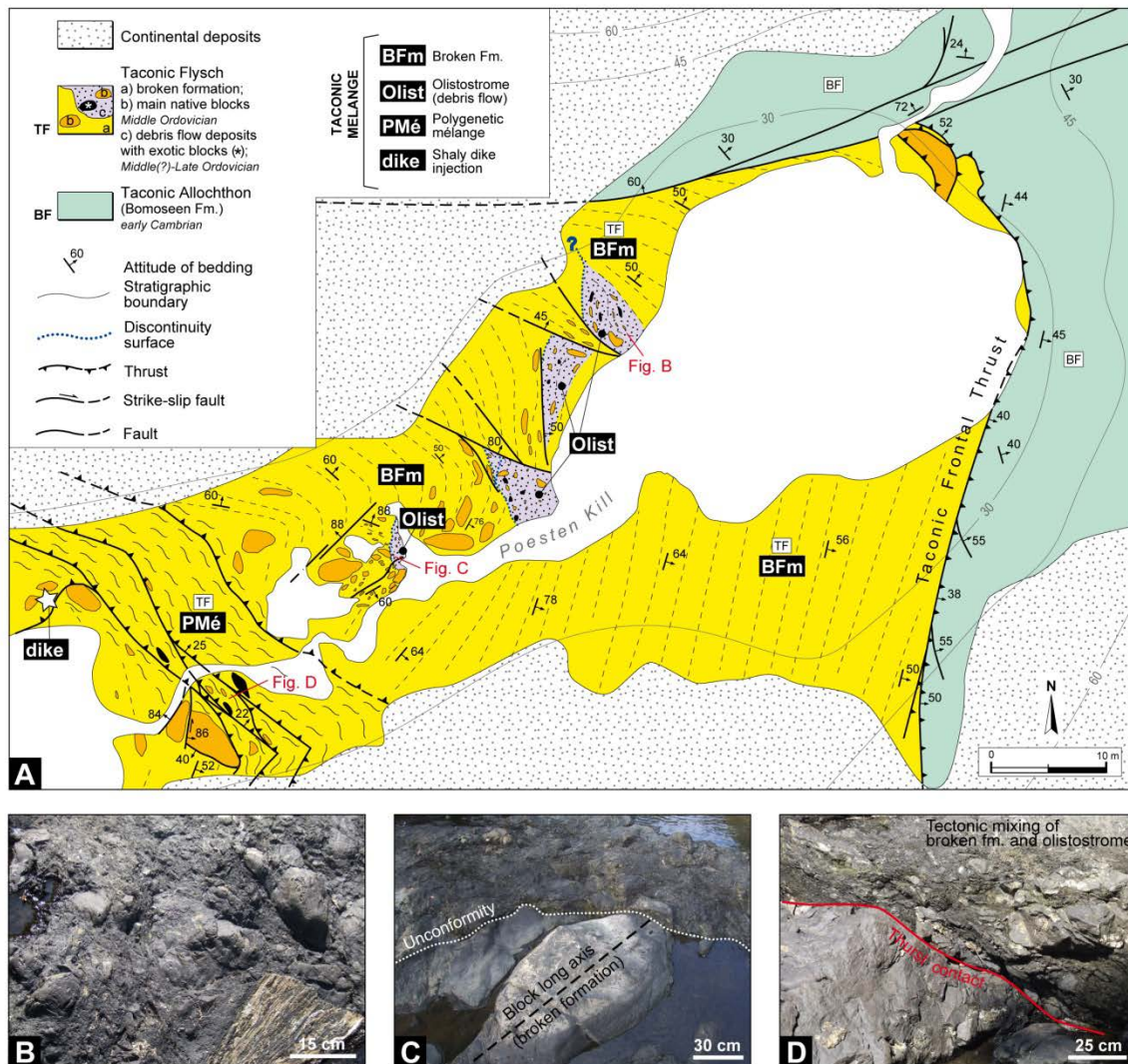


Fig. 11. (A) Geological map of the Poetsenkill Gorge in the Northern Appalachians (Troy, NY; modified from Festa et al., 2012a). Polygenetic mélange formation in this area was associated with thrust faulting during the Taconic orogeny (Taconic Frontal Thrust). The “structurally ordered” block-in-matrix fabric (SE corner of the map) of this mélange formed due to mixing of blocks from debris flow deposits (olistostromes) and broken formations. (B) Close-up view of a debris flow deposit with rounded clasts in a fine-grained matrix; (C) unconformable contact between the olistostrome and the broken formation; (D) close-up view of the polygenetic mélange (hanging wall) formed by thrusting-related mixing of broken formation and debris flow (olistostromes) deposits.

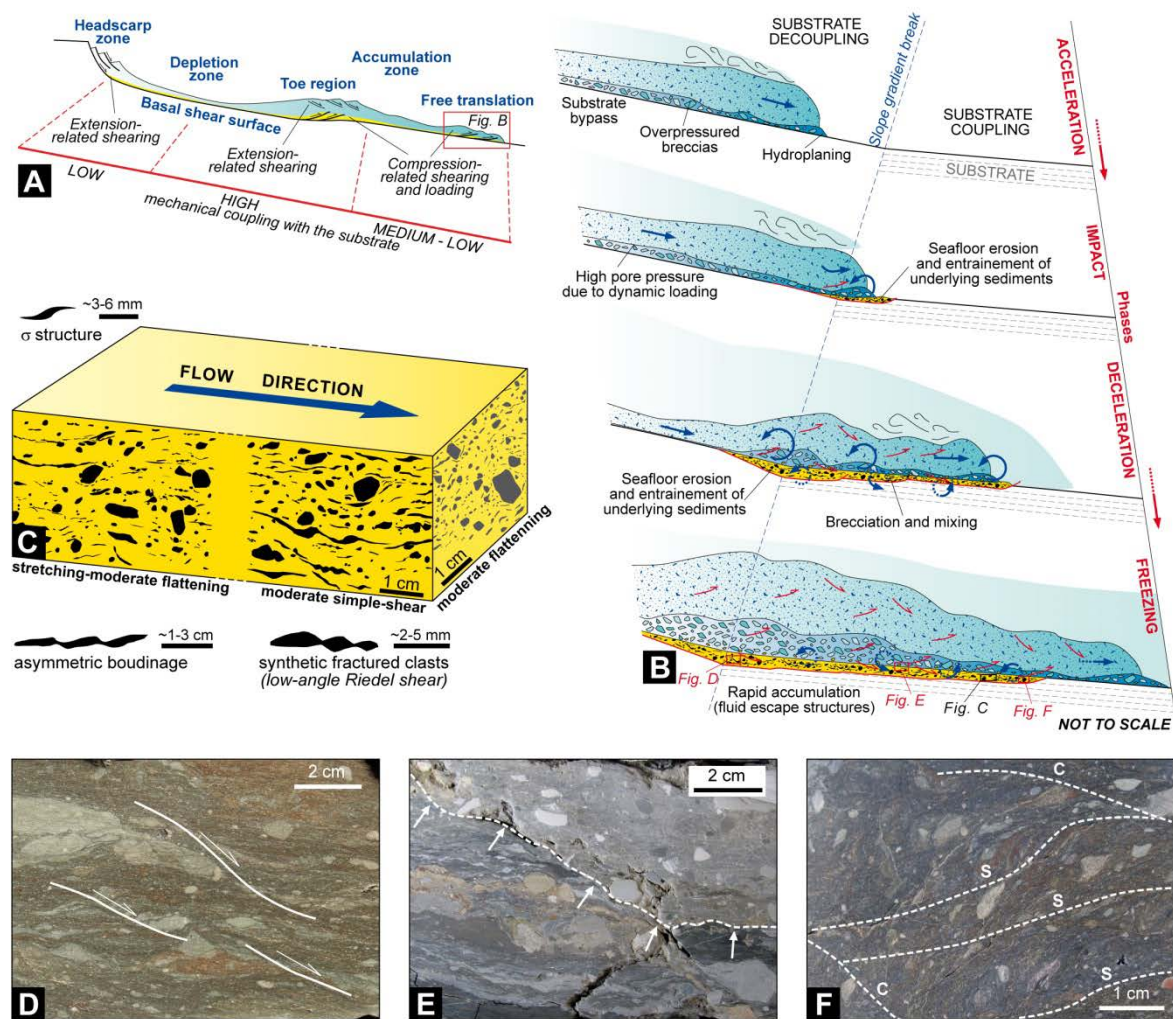


Fig. 12. Conceptual profile of downslope transformation of cohesive olistostromes. (A) Downslope evolution of a submarine landslide complex displaying distinct zones of mass transport-related deformation and deformation mechanisms (modified from Ogata et al., 2014a). Relatively thick, matrix-dominated portions and matrix-rich shear zones represent the products of such localized, enhanced progressive deformation (from extensional to compressional, and their combination), characterizing specific portions of the MTD. (B) Progressive evolution of different phases of a slide mass, involving various stages of acceleration to deposition, with emphasis on dynamic loading-related substrate erosion/bypass (modified from Ogata et al., 2014b). (C) Block diagram of an olistostromal body and its flow direction, showing simple shear deformation of clasts at its base (modified from Pini et al., 2012). Photographs of polished hand samples from the basal shear zone of olistostromes in the Northern Apennines (Italy): (D) Extensionally sheared layers show a planar anisotropy crosscut by low-angle extensional shear surface (R shear: white lines) (modified from Festa et al., 2013); (E) superposition along an erosive surface (white arrows and dashed line) of a brecciated lenticular body onto extensionally sheared shaly layers (modified from Festa et al., 2013); (F) contractional shear zone, showing the reorientation of elongated hard clasts to an S-C fabric (white dashed lines; modified from Festa et al., 2015c).



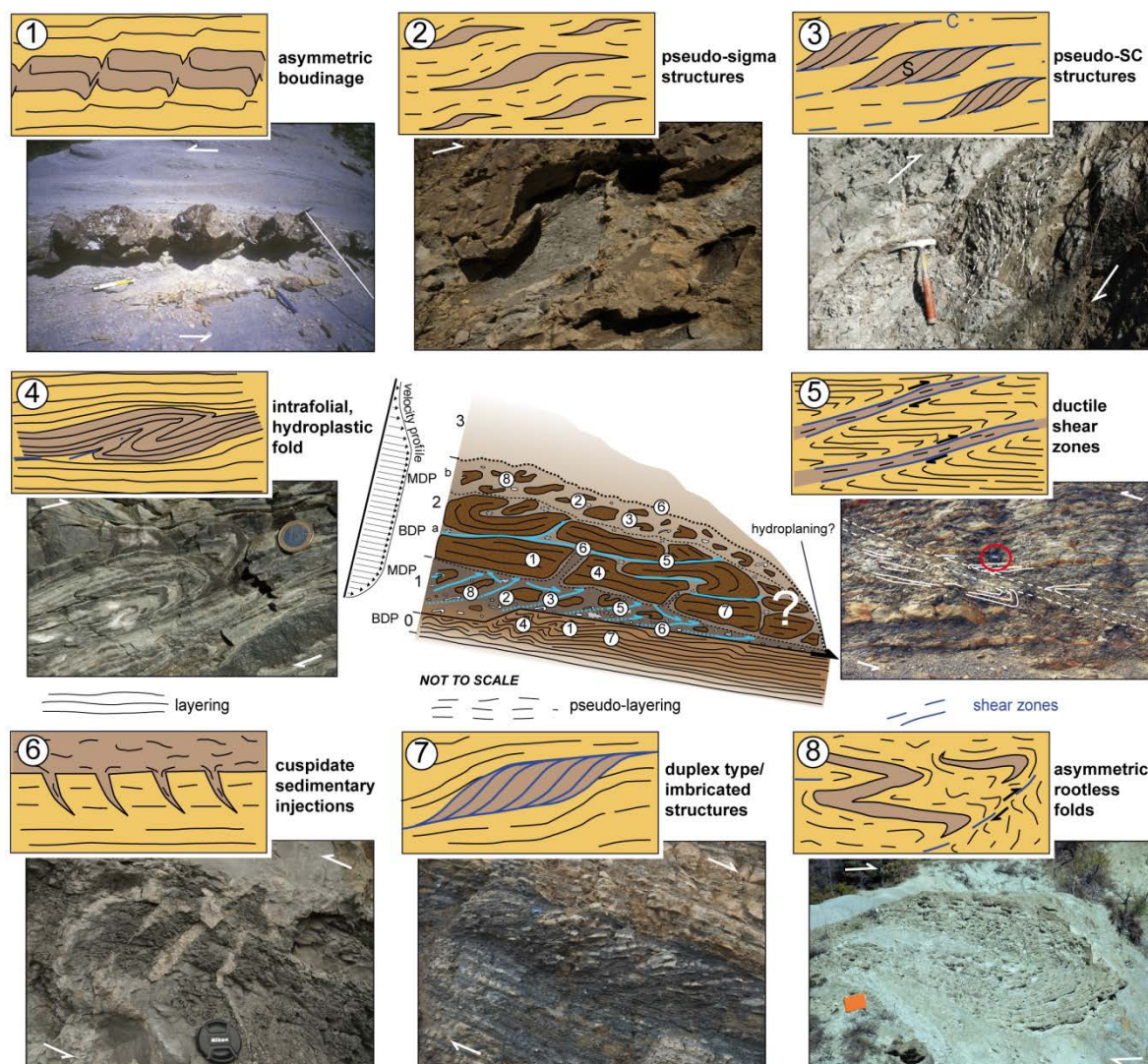


Fig. 13. Simplified cross-sections of ductile and brittle-ductile deformation fabrics and their kinematic indicators as discussed in the text. Field photographs depict some of the most representative examples of these structures. The cartoon with a velocity profile displays the main internal subdivisions, facies associations, and mesoscale structures of an evolving slide body (modified from [Ogata et al., 2014b, 2016](#)).



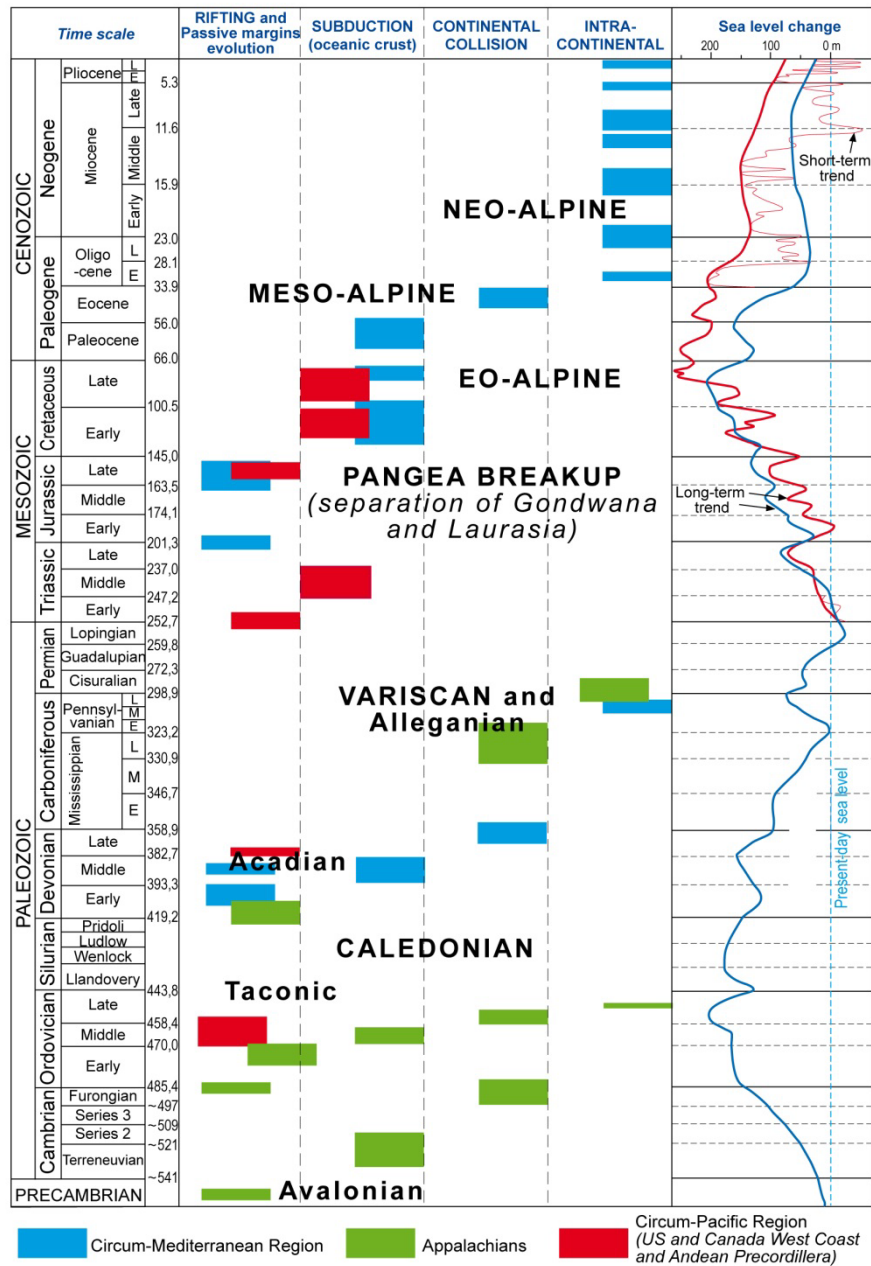
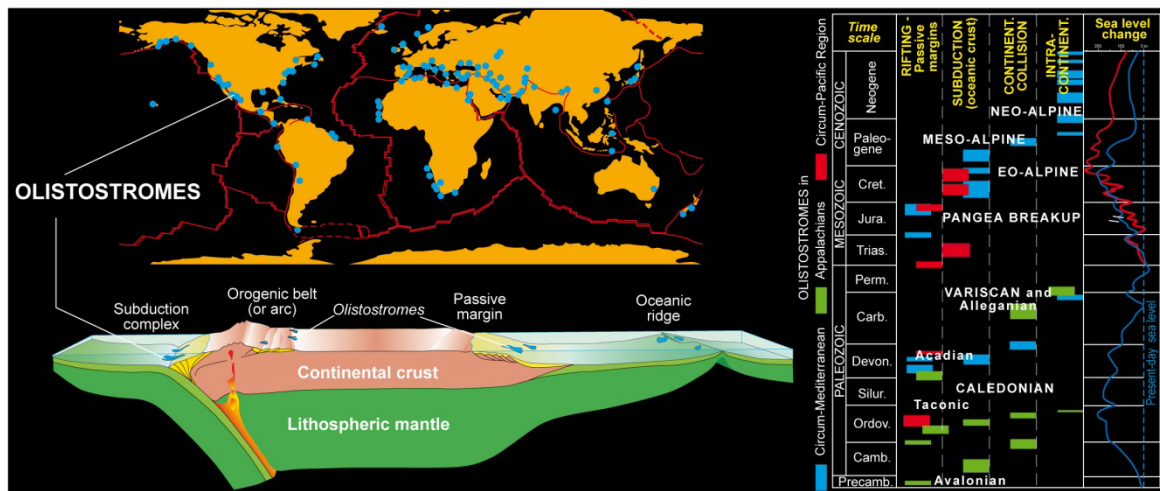


Fig. 14. Correlation between main orostromal events (cited in the text), tectonic settings, orogenic stages and sea-level changes in the Circum-Mediterranean Region, Appalachians and Circum-Pacific Region (the West Coast of the USA and Canada, and the Andean Precordillera). Sea-level changes from Haq et al. (1987) for the Cenozoic, and Haq and Schutter (2008) and Snedden and Liu (2010) for the Paleozoic and the Precambrian.

## Graphical Abstract



## Highlights

- Precambrian to Late Cenozoic olistostromal occurrences are compared worldwide.
- Olistostromes represent ancient examples of modern submarine MTDs and MTCs.
- Olistostromes are significant markers of tectonic and climatic events worldwide.